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GLACIAL GEOLOGY OF NORTH-CENTRAL KEEWATIN, NORTHWEST TERRITORIES, CANADA

BY R. S. TAYLOR

ABSTRACT

Much of the area is mantled by till, whose thickness exceeds 100 feet in places. A complex history of glacierization is recorded by drumlins of diverse patterns, by eskers, and by features eroded in bedrock. The eskers form a dendritic pattern across the area, and are believed the result of regressive deposition from consequent superglacial streams. Some transitory proglacial lakes existed, but there has been no post-Wisconsin marine submergence of the area.

The recorded sequence of ice movements through the area from oldest to youngest is (1) to southeast, (2) to north, (3) to N. 30° W., and (4) to N. 60° W.

A carbon-14 age determination of 4140 ± 150 years was made on peat buried by the last ice advance.

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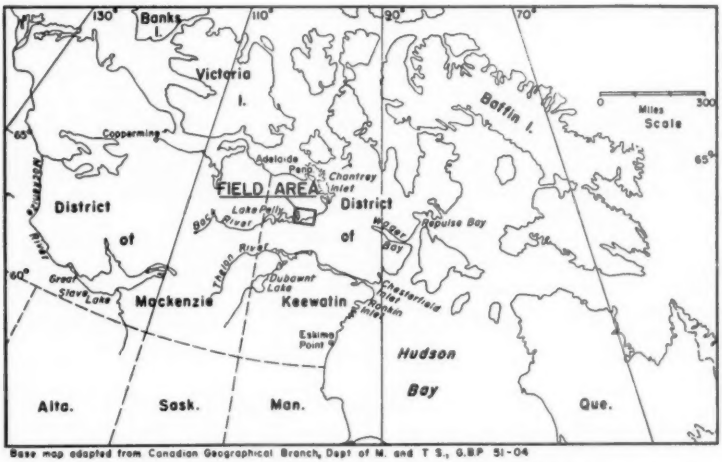
INTRODUCTION

A field party from the University of Minnesota supervised by Dr. W. J. Breckenridge, Director of the University's Museum of Natural History, worked in the Back River country (Fig. 1), in north-central Keewatin, N.W.T., from July 13 to August 6, 1953. As party geologist, the writer conducted a reconnaissance and collected samples for laboratory work. This paper constitutes a part of the results of his investigations (Taylor, 1955, Univ. Minn., unpublished M.S. thesis).

Back (1836; King, 1836) and Anderson (Clarke, 1940; 1941) took parties along the back River in the middle of the last century, and Schwatka traveled through part of the field area in 1879 (Gilder, 1881). The Upper

Back River Shoran Station was established within the field area in 1951 by the Geodetic Survey of Canada. None of these parties made detailed investigations, but Back, King, and the Office of the Quartermaster General, U. S. Army. The writer was aided by a grant in aid from the Penrose Bequest of The Geological Society of America.



Base map adapted from Canadian Geographical Branch, Dept of M. and T S., G.B.P. 51-04

FIGURE 1.—INDEX MAP OF NORTHEASTERN CANADA

and Anderson kept diaries on the regions traversed.

The eastern part of the area has a thin discontinuous cover of glacial drift on Precambrian granitic rocks that trend northeast. A narrow Proterozoic (?) quartzite syncline, trending northeast, lies just northwest of the base camp (Pl. 1). A till plain begins several miles west of the base camp and extends west at least to Lake Pelly.

ACKNOWLEDGMENTS

The writer is grateful for assistance from several branches of the Canadian Department of Mines and Technical Surveys and from the Northern Laboratories of the Defense Research Board of Canada, at Churchill, Manitoba. Appreciation is expressed to Dr. S. S. Goldich and especially to Dr. H. E. Wright, Jr., of the Department of Geology and Mineralogy of the University of Minnesota, and to Dr. W. J. Breckenridge, for valued assistance, criticism, and advice.

The expedition received financial support from the Wilkie Foundation, Minneapolis,

METHODS OF STUDY

Field Investigations

Small boat and aircraft enabled the writer to extend the work beyond the base-camp area and make a trip down the Back River from Lake Macdougall.

Starting from the Upper Back River Shoran Station elevations were determined throughout the area with a Paulin altimeter, and they are believed correct to within 50 feet.

Bearings were taken with a Brunton compass and corrected for declination (*see* maps of Canadian National Topographic series: Aberdeen Lake, N.W. 64/104 and Ogden Bay, N.W. 66/104, 1953 editions). Attempts to determine declination in the field yielded highly variable results, with declinations ranging from 8° W. to 11° E. in seven of nine attempts in the base-camp area. Three miles east-northeast of the base camp measured declination was 40° E. The indicated declination on the pertinent map is about 7° W.

Laboratory Investigations

Standard techniques were used for the pipette and wet sieve analyses of the minus-4-mm

fraction of the soil samples. Temperatures of sedimentation graduates fluctuated with room temperature, and this introduced certain errors (Robinson, 1915).

Aerial Photographs

Several hundred 9 by 9 inch trimetregon aerial photographs with a vertical scale of *ca.* 1/40,000, alternate prints along flight lines over the field area, were purchased from the Royal Canadian Air Force. Dr. V. K. Prest of the Geological Survey of Canada kindly arranged the loan of 80 additional prints from the Survey that permitted stereographic coverage of a strip about 20 miles wide along the south edge of the area covered by Plate 1. Most of the information presented on this map represents an extension of field data by photographic interpretation.

Glacial Erosion

Striae, Grooves, and Friction Cracks

Detailed observations were made on glacial erosion phenomena for their pertinence in deciphering the history of ice movements. Striae are found on many outcrops but least often on gneiss, which is affected most by frost action. Observations show three striae concentrations, as indicated by the rose diagrams on the map (Pl. 1), at about north, N. 30° W., and N. 60° W. In the base-camp area there is a fourth striae orientation at N. 90° W. Many outcrops have two sets of striae, and some have three. Deflection of striae caused by local topographic control is common, particularly along the Back River.

Grooves up to 6 feet wide were observed on outcrops of gneiss, syenite, and quartzite. Orientation of most grooves is close to north. Striae that cross grooves may indicate the absolute direction of movement of the striae-forming ice: as the ice moves out of the groove basal pressure is increased and greater abrasion occurs (Fig. 2). For example, at station 14 a groove oriented N. 7° W. has been crossed by ice moving N. 42° W. and N. 65° W.

Friction cracks occur in quartzite and are similar to the type Harris (1943, p. 249) termed crescentic fractures, in which the horns point in the direction of movement. The 3-to-4-inch-wide chip between the horns was missing in many instances. Where the fractures were aligned close together a rough-bottomed groove developed (Fig. 3), a probable forerunner of the normal smooth-surfaced groove.

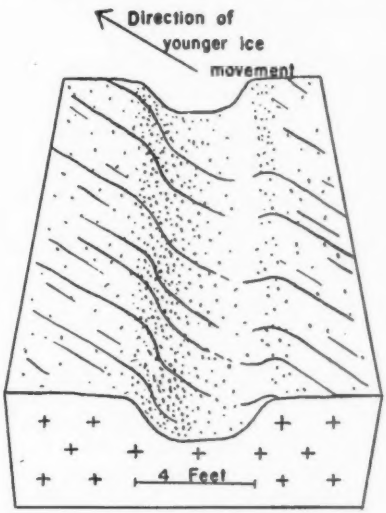


Figure 2.—Striae Crossing Groove
Diagrammatic sketch indicating direction of ice movement.

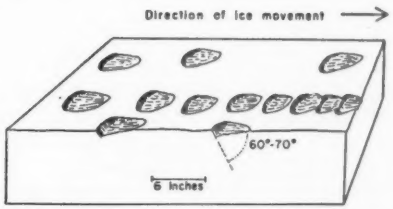


Figure 3.—Crescentic Fractures in Quartzite
Diagrammatic sketch.

Roches Moutonnées

Most of the bedrock hills of the area are asymmetrical because of scouring of the stoss sides and plucking on the lee. Steep slopes face north and west except on the southeast limb of the quartzite syncline adjacent to the base camp, where the northwesterly dip of the beds controls the ridge form. Striae which cross small *roches moutonnées* may permit determination of the absolute direction of movement of the younger ice in a manner

analogous to that used with grooves. At station 12 a *roche moutonnée* oriented N. 12° W. has been crossed by ice moving N. 57° W.

Till Grooving

Parallel light-gray streaks noted on the aerial photographs of the northern part of the field area proved to be shallow grooves in till, 60–100 feet wide and less than 20 feet deep. The light-colored grooves contained green vegetation while the sparse growth of the drier slopes was brown. In the east the till grooves were oriented N. 38° W.; in the west, N. 33° W.

Glacial Deposition

General

The extent of glacial drift, as mapped from aerial photographs, is given on Plate 1. Less than one-third of the eastern section of the area is covered by drift. In the western section, the till plain, drift covers more than two-thirds of the surface. Between these sections is a narrow zone with one- to two-thirds of its surface covered by drift.

Drift deposition has diminished local relief. The preglacial drainage pattern and most of the preglacial topography west of the junction of the Back and Meadowbank rivers has been obscured.

The maximum observed thickness of till occurs between stations 70 and 71, where the Back River flows in drift between drift banks 90–100 feet high. No buried soil horizon or other differentiation between tills was seen in any exposure.

Crag-and-tail features occur in the southeast; the tails stream out N. 32° W. for about a mile from outcrops of a dike. Drumlins, eskers, and related outwash deposits are numerous, and the absence of recognizable moraines is puzzling. There are several anomalous areas whose exact nature and origin is uncertain.

Drumlins

While not the largest features, drumlins dominate the landscape of the till plain, forming an apparently unending succession of low rolling hills with intervening cotton-grass meadows. They are commonly ovoid in plan view, with lengths to more than 4 miles and

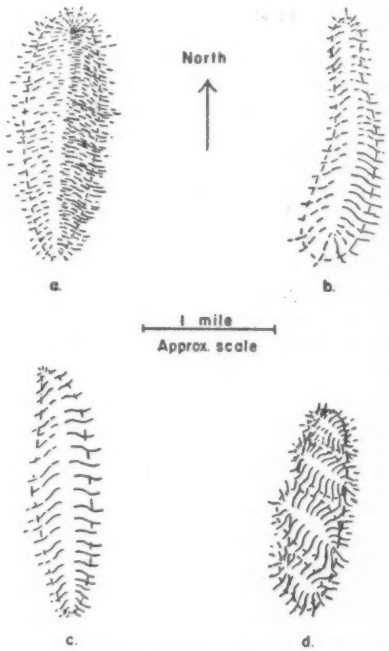


Figure 4.—Drumlins
(a) Relatively undeformed. (b), (c), and (d) are north-south drumlins deformed by ice moving from the southeast. Generalized sketches from aerial photographs.

length:width ratios from about 6:1 to 2:1. Some are higher than 100 feet, with side and end slopes no more than 14° and 5° respectively.

The orientations of several hundred undeformed drumlins, determined from aerial photographs and maps, are plotted on Plate 1 in rose diagrams. The north-south drumlin set largely controls the lineation of the lakes and minor watercourses in the till plain. Three other drumlin sets are apparent in the rose diagrams, oriented about N. 30° W., N. 60° W., and N. 45° E. This method of plotting tends to emphasize the minor patterns because the north-south pattern has the highest percentage of deformed drumlins. The deformation has many forms (Fig. 4) and has been caused by

ice moving from the southeast obliquely across the drumlins. Lee (Personal communication; Lord, 1953, p. 2) investigated similar deformations in southern Keewatin, and Bird (1953, p. 222) has reported superimposed drumlin patterns in central Keewatin.

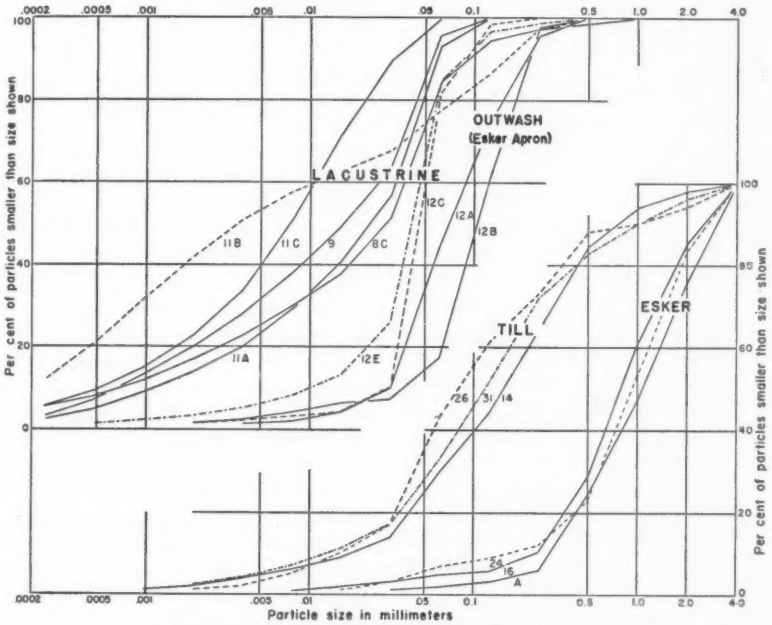


FIGURE 5.—TEXTURAL ANALYSES OF SOIL SAMPLES

Sample no.	Station no.	Sample depth	Remarks	Sample no.	Station no.	Sample depth	Remarks
A	Base camp	42 inches	Stratified	12 B	19	45 inches	
8 C	14 A	5 inches	Over stratified sands	12 C	19	90 inches	
				12 E	19	11 feet	
9	14 B	0–4 feet	Channel sample	14	52 A	8 feet	
11 A	18	—	40 ft. section; bottom unit	16	56	4 feet	No stratification
				24	68	5 feet	Stratified
11 B	18	—	Massive cntr. unit	26	70	20 feet	In river bank
11 C	18	—	Massive upper unit	31	85	—	40 ft. above river
12 A	19	15 inches	*See* Fig. 6				

The origin of drumlins is a controversial subject. Gravenor (1953, p. 674–681) summarized erosional and depositional theories, and advanced a modified erosional theory. In the area under consideration drumlins developed by both erosion and deposition. A depositional type is illustrated by the drumlin of granitic till at station 98, on the southeast limb of the quartzite syncline; it is the result of deposition from locally overloaded basal ice as the ice thinned while rising 250 feet out of the Back River valley. An erosional type is illustrated by the relatively small drumlins with a northwest orientation, which are formed of till eroded from the larger, pre-existing, north-south drumlins (Fig. 4d).

The till is a stony silty sand. Textural analyses of three grab samples collected at widely separated locations along the Back River are given in Figure 5. The small amount of clay-sized particles present in seven samples taken in drumlins (average: 3 per cent) support the contention of Gravenor (1953, p. 676–677), as opposed to Wilson (1938, p. 42–45), that clay is not necessary to drumlin formation.

The pink color (Munsell 10R 8/2) of the

western till reflects the quantity of bright pink potash feldspar (mostly microcline) and of clear and rose quartz grains. Eastward the feldspars are gray, as is the till (2.5Y 8/2).

Eskers

Eskers, rising 250 feet or more above the adjacent till plain, generally have broad crests but in places are marked by multiple sinuous ridges with kettle holes along them. Some eskers can be traced, with minor breaks, across the entire area (Pl. 1). In places they are composed of a succession of poorly defined hills. Ridge continuity is also broken by the occurrence of esker deltas, which are local expansions of eskers that form irregularly pitted and hummocky areas of considerable local relief and local radiate structure. Eskers of classic embankment cross section also occur, as at station 68. This esker, 50–75 feet high, 30–50 feet across the top, and with a base about 200–250 feet broad, was traced 5 miles to the south and 4 miles to the north into the much larger and commoner type.

Eskers and esker deltas are the most prominent features of the till plain, as they are over broad areas of Arctic Canada. The similarity of Mt. Pelly, which stands 500–600 feet above the surrounding terrain on Victoria Island (Washburn, 1947, p. 53–55), to certain deltaic deposits has been noted by Bird (1951, p. 27) and Lougee (Jenness, 1952, p. 947). The writer believes that Mt. Pelly is a very large esker delta, similar to some of those in the middle Back River region.

The eskers have a dendritic pattern created by streams flowing to the northwest and north. The pattern crosses the Back River valley, disregarding major topographic obstructions. Eskers rise as much as 400 feet downstream in a distance of 12 miles. Tyrrell (1896, p. 186F), Wilson (1939, p. 122–127), and Fortier (1949, p. 22) commented on the similar characteristics of eskers elsewhere in northern Canada.

Flint (1947, p. 151–152) attributes most eskers to subglacial streams principally on the evidence from eskers that follow topographic depressions. The surface melt of thin and stagnant or nearly stagnant ice works downward through various openings and at the base of the ice enlarges a system of openings to form tunnels. The lowest possible channelways (valleys) are sought, and the water flows, chiefly under hydrostatic pressure, into some body of ponded water or perhaps into the open.

The hypothesis of subglacial origin does not appear to apply to most eskers in the field area, for the following reasons.

(1) The eskers do not follow the lowest possible channelways. In the north they swing around parallel to the regional slope toward the Arctic coast (Bird, 1953, p. 223), but they cross the second regional slope that is followed east-northeasterly by the Back River. On the Back River lowland eskers locally parallel the stream, but then rise obliquely as much as 400 feet out of the valley.

(2) If wasting ice were thoroughly penetrated by meltwater it would be relatively warm. A warm glacier 200–250 feet thick (Flint, 1947, p. 17) would flow at the base, preventing the development of openings and tunnels. Yet some eskers in the field area are 200–250 feet high.

(3) Stagnant or near-stagnant ice could not so confine subglacial meltwater as to permit a hydrostatic head to develop that would be capable (a) of building eskers to a height of 250 feet, (b) of propelling a loaded stream up a grade of 33 feet/mile for a distance of 12 miles, or (c) of creating and continuously extending headward a dendritic subglacial stream pattern whose length can be measured in tens or scores of miles.

The dendritic pattern of eskers is believed to be the result of development of an integrated consequent drainage pattern upon the waning ice sheet as a moderating climate permitted more or less continuous melting. Wilson (1939, p. 127) has clearly expressed this concept. The topography of the surface of a waning ice sheet would probably reflect underlying major topographic irregularities, particularly near its edge, so many of the consequent streams would eventually be superimposed on bedrock valleys. The eskers resulted from regressive (headward) deposition by superimposed streams flowing in ice-walled gorges at the edge of the wasting ice. The material for the eskers was supplied by shear planes such as those reported by Goldthwait

(1951, p. 569–571) marginal to the Barnes Ice-cap on Baffin Island, or similar ones observed by the author at heights to 350 feet above the toe of the Greenland ice sheet near Thule. Additional material could be supplied by englacial debris in the lower part of the ice. Formation of eskers in this fashion and particularly of esker deltas would proceed most favorably in proglacial lakes.

The classic- or normal-type esker near station 68 may be of subglacial origin. It grades in both directions into the larger and commoner variety. Other such esker segments were observed on aerial photographs of the area, but they are relatively short. Nothing in the above discussion precludes the develop- of relatively short segments of subglacial eskers in favorable locales under the edge of the wasting ice.

Most of the esker material is clean bouldery sand. Stratification was observed in all pits dug in the sides of the base-camp esker, with dips to the northwest or away from the adjacent esker ridge. Widespread outwash deposits (esker aprons) of gravel, sand, and fines are commonly associated with the flanks of the eskers. Where an esker is discontinuous there is generally a low-lying outwash deposit on the till plain between adjacent esker segments.

Representative textural analyses of esker and esker apron samples, taken over a 40-mile east-west span, are given in Figure 5.

Anomalous Areas

A number of areas were outlined on Plate 1 from aerial photographs as anomalous. These areas are hummocky and topographically confused, locally exhibit braided channeling, and contain small randomly oriented lakes in contrast to the general north-south lineation of lakes in adjacent areas. The sides of some drumlins and hummocks appear undercut by stream action. Eskers occur in some of these areas.

The origin of the mapped anomalous areas has not been determined. Bird (1953, p. 221–222) mentions small abandoned overflow channels in sandy areas between the Back River lakes and the Thelon River lakes, and describes water-eroded channels in the till occupied in part by multiple-ridge eskers. Some areas may be of hummocky moraine and/or dead-ice deposits (Bird, 1953, p. 222) that have been dissected by meltwater streams. Those in the southwest part of the mapped area may be related to overflow spillways of a proglacial lake occupying the Thelon basin (Bird, 1953, p. 227; 1954, p. 459). In the west center and center they may best be tentatively classed as superglacial stream valleys superimposed on the land surface upon or near drainage divides.

Proglacial Lakes

Evidence

Two lines of evidence indicate the existence of proglacial lakes in the middle Back River region during late-Wisconsin time. One is the occurrence of many esker deltas in the Back River basin. The deltas which occur on the Arctic slope within the mapped area at elevations of about 450 feet suggest deposition in an expanded Arctic Ocean, but there is no evidence elsewhere in the area for such high marine limit. Bird (1954, Fig. 1) indicates that the marine limit lies well to the north of the Back River-Arctic Ocean divide. Perhaps these land forms were created without the presence of a body of water, but the mechanism by which this could occur is not understood.

The second line of evidence for proglacial lakes lies in the nature and stratigraphic position of samples collected at stations 14A, 14B, 18, and 19 (Fig. 5), near the base-camp esker. In places lake sediments mantle outwash of the esker; other more massive lake deposits, with their bases concealed, lie alongside this esker, are not covered by outwash, and are believed younger than the esker.

The station 19 exposure (Fig. 6) is important to an understanding of the late-glacial history of the area. Its interpretation is as follows: as the edge of the penultimate ice withdrew to the southeast, a proglacial lake covered this area, and silt (sample 12E) was deposited. As the ice cleared the Back River valley and the proglacial lake drained, a lesser lake was formed by a dam of glacial drift at station 15. The several peat layers represent fluctuations in the level of this lake, with the uppermost

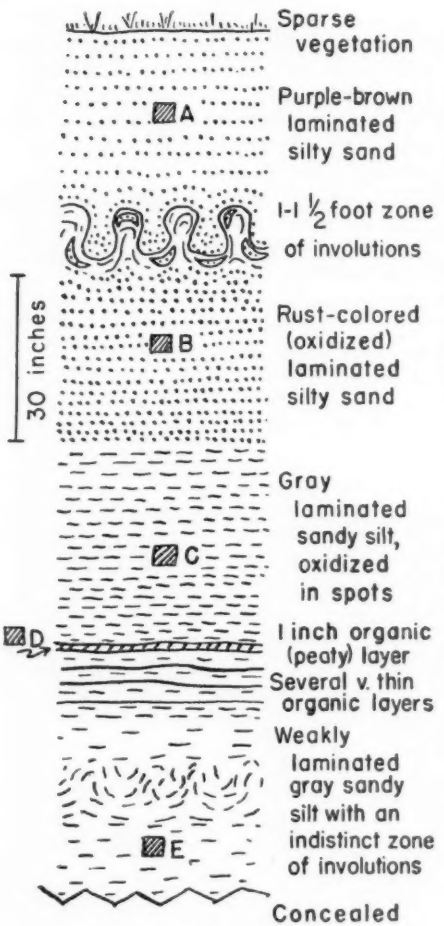


FIGURE 6.—STATION 19 EXPOSURE

Site of sample series 12A–E, a river-cut bank actively caving. Section starts about 25 feet above the river.

layer of vegetation (12D) accumulating after the level of the lake had been lowered by partial erosion of the drift dam. Readvance of the ice from the southeast produced a younger proglacial lake in which the upper silt unit (12C) was deposited. This ice then overran the lake and continued northwestward beyond the area. In its retreat it left the base-camp esker, several hundred yeards distant, whose flanking sands (12B, A) cover the lake beds. Subsequent erosion by the Back River has dissected the deposits to a depth of more than 30 feet.

The involutions in this exposure may be the products of frost action (Sharp, 1942, p. 113–133). Their symmetry precludes ice shove; there is no disturbance of the upper layers by glacier overriding; the present overburden is inadequate to have produced them; and the uppermost zone was formed after the withdrawal of the last ice sheet according to the stratigraphic evidence, thus eliminating the weight of the ice as a factor. While the last argument is not applicable to the lower involutions, they may have had a mode of origin similar at an earlier time.

Extent and Duration of Proglacial Lakes

At station 1 the base-camp esker rises 240 feet above river level. At the time of its formation, the Back River valley was blocked by the ice front to a height at least equal to this, and a proglacial lake probably extended to the foot of the outlet of Lake Macdougall. At an earlier stage, the basin may have been blocked to a greater depth. If these lakes had spillways to the north, they were not identified in the field area—unless they are related to the anomalous areas.

All the evidence on duration is negative, for no shore-line features were identified. The lakes were apparently of a transitory nature, indicating that the ice front receded rapidly.

MARINE SUBMERGENCE

There is no evidence of post-Wisconsin marine submergence in the field area nor for some distance northward. Heights checked in the area range from 100 feet at the head of the rapids near station 47 to 600 feet on the divide north of the base camp. With respect to the field area, Bird's (1954, Fig. 1) conclusion of no post-Wisconsin marine submergence is upheld.

Jenness (1952, p. 941–942) suggests an average emergence of the western part of the Archipelago of perhaps 400–500 feet. Bird (1953, p. 229) has reported marine features as high as 360 feet in the middle of the Thelon basin. In view of these data, a reason was

sought to account for the lack of similar recent marine submergence of the Back River basin.

It is generally held that the emergence of the lands south of the Back-Thelon divide is caused by isostatic recovery from ice load. The elevation of the lands to the north, however, may result from regional unwarping of the crust; Washburn (1947, p. 59) mentioned this possibility. It is possible, then, that the Back River area is part of a structurally positive unit of the crust acting as a hinge block for rising crustal segments both to the north and south; the southern movement being isostatic, the northern movement being epeirogenic in whole or major part. If this is correct, the lack of recent marine submergence would be attributable to a lesser depression of the area because of the regional structure.

Ice Movements

The principal striae sets, in order of decreasing size and abundance, are oriented N. 60° W., N. 30° W., and north. Field evidence indicates that this order also represents increasing age, and that in all three the ice movement was toward the northern quadrants. The older striae have been partly destroyed by younger ice movements and frost action. In addition to these three sets, an east-west set was found only on the left bank of the Back River just east of the base camp. Its orientation probably reflects topographic deflection of the youngest ice, diverted westward from its general course to N. 60° W.

The principal drumlin sets, in order of decreasing abundance, are north, N. 30° W., and N. 60° W., the reverse of the order for striae. This order represents decreasing age. The oldest drumlins are larger and very compact and resisted reorientation by younger ice, which may have been thinner and shorter-lived. The rose diagrams of drumlins indicate that neither of the two younger movements had much effect on the ill in the northern part of the field area, although the eskers show that the youngest ice extended over it.

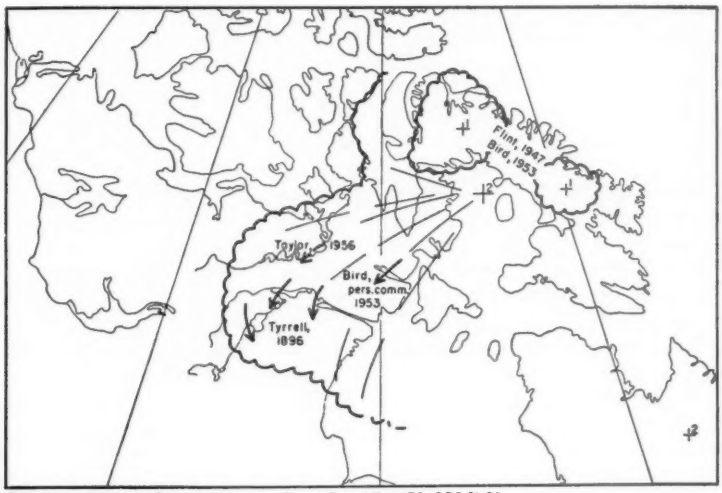
In addition to these three major drumlin sets, there is a minor set oriented northeast. Neither the age nor absolute direction of movement of the ice creating the northeast-trending drumlin set was determined in the field. Field evidence from Wager Bay, 200 miles east of this area, indicates movement to the southwest (Bird, Personal communication). This evidence, the lack of any definite bedrock record in the field area, and climatological requirements support the opinion that ice movement to the southwest is the oldest of record within the mapped area.

The combined evidence on direction of movement gives the following order: (1) to southwest, (2) to north, (3) to N. 30° W., and (4) to N. 60° W.

Figures 7–10 depict several stages in the Wisconsin glacierization of northeastern North America. Climatological requirements of an initial source area suggest early growth of glaciers on Baffin Island (Flint, 1947, p. 233; Bird, 1953, p. 217), followed by their confluence and extension southwestward across the Wager Bay area, the middle Back River region, and the Thelon basin (Fig. 7). Downie *et al.*, (1953) describe an ice movement west of Hudson Bay which fanned out to south and west (Fig. 8), south of a disturbed zone between Eskimo Point and Coppermine. They also locate a large terminal moraine immediately east of the Mackenzie River delta. The writer suggests that this is an end moraine of ice which invaded Victoria and Banks islands from the mainland to the south and southeast (Washburn, 1947, p. 49, 59; Jenness, 1952, p. 950). The disturbed zone (Fig. 9) may be correlated in part with Wilson's master esker (1939, Fig. 2A-B, p. 124), which he suggested occupied a position between two ice sheets (lobes?) one centered near Lake Dubawnt and the other on the Back River. Downie *et al.*, (1953) state that north of the disturbed zone the flow, probably formed by another ice sheet (lobe?) swept from Hudson Bay to the Arctic Coast.

The time of maximum glaciation of the Back River region is depicted in Figure 9. The ice apparently continued northward and formed the arcuate drumlin belt from Adelaide Peninsula to eastern Victoria Island (Fortier, 1948, p. 7–8; Jenness, 1952, p. 943), perhaps extending to northeastern Banks Island (Jenness, 1952, p. 950).

As Wisconsin glacierization waned a residual



Base maps adapted from Canadian Geographical Branch, Dept of M and T S, G B P 51-04

FIGURE 7.—INCEPTION OF WISCONSIN GLACIERIZATION; COALESCENCE AND EXPANSION OF PIEDMONT GLACIERS

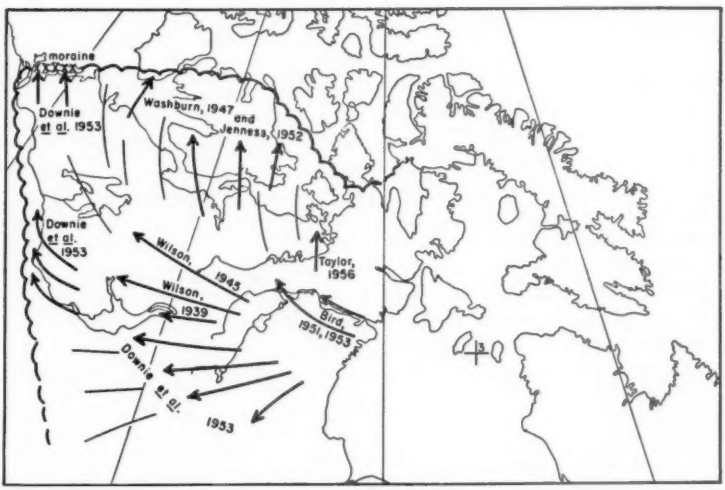


FIGURE 8.—MAXIMUM WESTERLY EXTENT OF WISCONSIN ICE FROM HUDSON BAY AREA

FIGURES 7–10.—POSTULATED PA

Heavy arrows and accompanying names indicate areas of, and references to, ice movement. centers of distribution, similar numbers being contemporaneous. Indicated movements are not ne

ice sheet (Flint, 1947, p. 242–243) was left on the uplands west of Hudson Bay and north of Latitude 60° (Fig. 10). Bird (1954, p. 459–460) suggested this, and Lee (Lord, 1953, Fig. 1) mapped what appears to be a shifting center of distribution in south-central Keewatin. Perhaps the residual ice sheet broke into several distinct upland glaciers as suggested by Tyrrell (1896, p. 179F) for his remnant of the Keewatin glacier. It is also possible that the center of flow migrated northward along a path 140–160 miles inland from the present

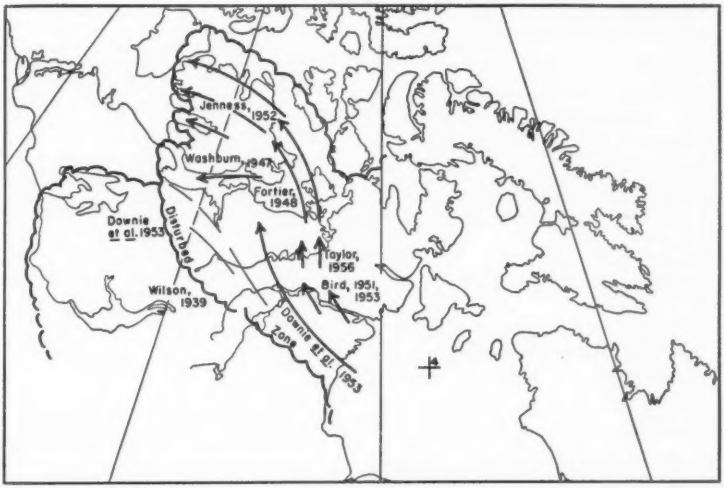


FIGURE 9.—GREATEST NORTHWARD EXTENSION OF WISCONSIN ICE FROM HUDSON BAY AREA

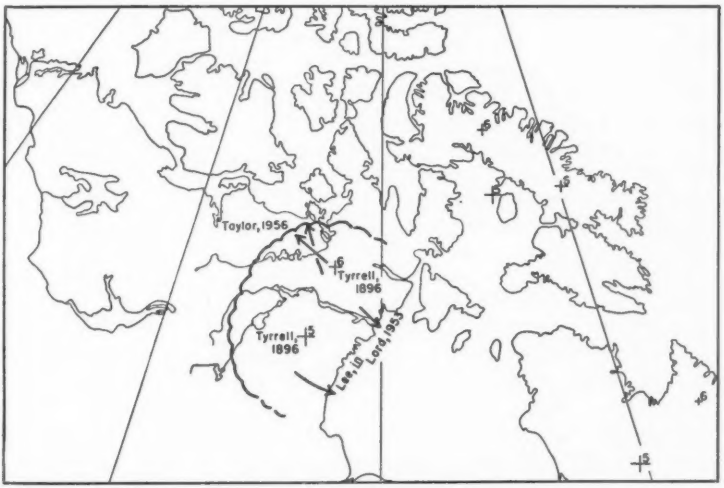


FIGURE 10.—RESIDUAL WISCONSIN ICE SHEET, WEST OF HUDSON BAY

ISTORY OF WISCONSIN GLACIERIZATION

ces the arrows are inferred from texts. Light lines are conventional flow lines. Crosses are postulated ily immediately sequential. *See* Figure 1 for place names.

Hudson Bay coast. Such a migration could be correlated with the northward shift of the circumpolar storm tracks and zonal westerlies as the climate moderated (Willett, 1949, p. 43, 50). The end product, as recorded in the field area by the writer and at the mouth of Chesterfield Inlet by Lee, was an icecap terminating on the upland between Wager Bay and the Thelon-Back divide. From this cap, ice flowed N. 60° W. across the middle Back River region and southeast toward Chesterfield Inlet.

In Washburn's summary of pre-1942 findings on striae in the Canadian Arctic (1947, Pl. 3, p. 104–116) there are two contradictions to the ice movements plotted in Figures 7–10: Mathiassen's south-southeast-moving ice in Repulse Bay and Washburn's south-moving ice in southern Victoria Island. Both of these may be attributed to ice radiating from residual sheets on nearby uplands (Washburn, 1947, p. 60).

Carbon-14 Date[1]

The peat sample (12D) taken at station 19 (Fig. 6) has been dated by carbon-14 analysis at the Yale Geochronometric Laboratory (Preston *et al.* 1955, p. 959, sample Y-261): the determined age is 4140 ± 150 years.

This peat was buried during the last advance of ice through the field area; the uppermost overlying sands can be traced directly into the flank of the near-by base-camp esker. The radiocarbon age for this ice advance is surprisingly low; it is generally thought that any such expansion of the ice sheet would be somewhat older. Processes of weathering and organic activity may have produced undetected contamination, but, as Hunt points out (1955, p. 246), such contamination would be at a minimum in cold latitudes. The age is not inconsistent, however, with the concept of northward-shifting centers of distribution with recurrent but lessening advances of ice. This suggestion is supported by the age (5220 ± 340 years) indicated for an ice advance at Rankin Inlet (rePston *et al.*, 1955, p. 959, sample Y-231) and by the 6400-to-6700-year age indicated for the retreat of the ice after the Cochrane readvance (Preston *et al.*, 1955, p. 957, sample Y-222; Rubin and Suess, 1955, p. 485, sample W-136). The relationship of these ice movements to the Postglacial Thermal Maximum centered around 5000 years ago (Flint, 1953, Pl. 3) remains a problem.

References Cited

Back, George, 1836, Narrative of the arctic land expedition—in the years 1833, 1834, and 1835: London, John Murray, 663 p.

Bird, J. B., 1951, The physiography of the middle and lower Thelon basin: Canada Dept. Mines and Tech. Surveys, Geog. Br., Geog. Bull. 1, p. 14–29

—— 1953, The glaciation of central Keewatin, Northwest Territories, Canada: Am. Jour. Sci., v. 251, p. 215–230

—— 1954, Postglacial marine submergence in central Arctic Canada: Geol. Soc. America Bull., v. 65, p. 457–464

Clarke, C. H. D., 1940, 1941, Chief Factor James Anderson's Back River journal of 1855: Canadian Field Naturalist, v. 54, p. 63–67, 84–89, 107–109, 125–126, 134–136; v. 55, p. 9–11, 21–26, 38–44

Downie, M. J., Evans, A. G., and Wilson, J. T., 1953, Glacial features between the Mackenzie River and Hudson Bay plotted from air photographs: Geol. Soc. America Bull., v. 64, p. 1413–1414

Flint, R. F., 1947, Glacial geology and the Pleistocene epoch: N. Y., John Wiley and Sons, Inc., 589 p.

—— 1953, Probable Wisconsin substages and Late-Wisconsin events in northeastern United States and southeastern Canada: Geol. Soc. America Bull., v. 64, p. 897–920

Fortier, Y. O., 1948, Flights in 1947 over the regions of the magnetic pole and the mainland between the Arctic Coast, Great Slave Lake, and Hudson Bay, Northwest Territories: Canada Geol. Survey Paper 48–23, 11 p.

—— 1949, Indin Lake map-area (east half), Northwest Territories: Canada Geol. Survey Paper 49-10, 28 p.

Gilder, W. H., 1881, Schwatka's search: London, Samson Low, Marston, Searle, and Rivington, 316 p.

Goldthwait, R. P., 1951, Development of end moraines in east-central Baffin Island: Jour. Geology, v. 59, p. 567–577

Gravenor, C. P., 1953, The origin of drumlins: Am. Jour. Sci., v. 251, p. 674–681

Harris, S. E., Jr., 1943, Friction cracks and the direction of glacial movement: Jour. Geology, v. 51, p. 244–258

Hunt, C. B., 1955, Radiocarbon dating in the light of stratigraphy and weathering processes: Sci. Monthly, v. 81, p. 240–247

Jenness, J. L., 1952, Problem of glaciation in the western islands of Arctic Canada: Geol. Soc. America Bull., v. 63, p. 939–952

King, Richard, 1836, Narrative of a journey to the shores of the Arctic Ocean: London, R. Bentley, (2 v.) 312 and 321 p.

Lord, C. S., 1953, Geological notes on southern District of Keewatin, Northwest Territories: Canada Geol. Survey Paper 53–22, 11 p.

Preston, R. S., Person, Elaine, and Deevey, E. S., 1955, Yale natural radiocarbon measurements II: Science, v. 122, p. 954–960

Robinson, G. W., 1915, Note on the effect of changes in the viscosity of water on the results of mechanical analysis conducted at varying temperatures: Jour. Agr. Sci., v. 7, pt. 2, p. 142–143

Rubin, Meyer, and Suess, H. E., 1955, U. S. Geological Survey radiocarbon dates II: Science, v. 121, p. 481–488

Sharp, R. P., 1942, Pereglacial involutions in northeastern Illinois: Jour. Geology, v. 50, p. 113–133

Tyrrell, J. B., 1896, Report on the Doobaunt, Kazan, and Ferguson rivers and the northwest

[1] This information became available after first submission of the manuscript.

coast of Hudson Bay: Canada Geol. Survey Ann. Rpt., new ser., v. 9, pt. F, 218 p.

Washburn, A. L., 1947, Reconnaissance geology of portions of Victoria Island and adjacent regions, Arctic Alaska: Geol. Soc. America Mem. 22, 142 p.

Willett, H. C., 1949, Long period fluctuations of the general circulation of the atmosphere: Jour. Meteorology, v. 6, p. 34–50

Wilson, J. T., 1938, Drumlins of south-west Nova Scotia: Royal. Soc. Canada Trans., 3d ser., v. 32, p. 41–47

—— 1939, Eskers north-east of Great Slave Lake: Royal Soc. Canada Trans., 3d ser., v. 33, p. 119–129

—— 1945, Further eskers north of Great Slave Lake: Royal Soc. Canada Trans., 3d ser., v. 39, p. 151–153

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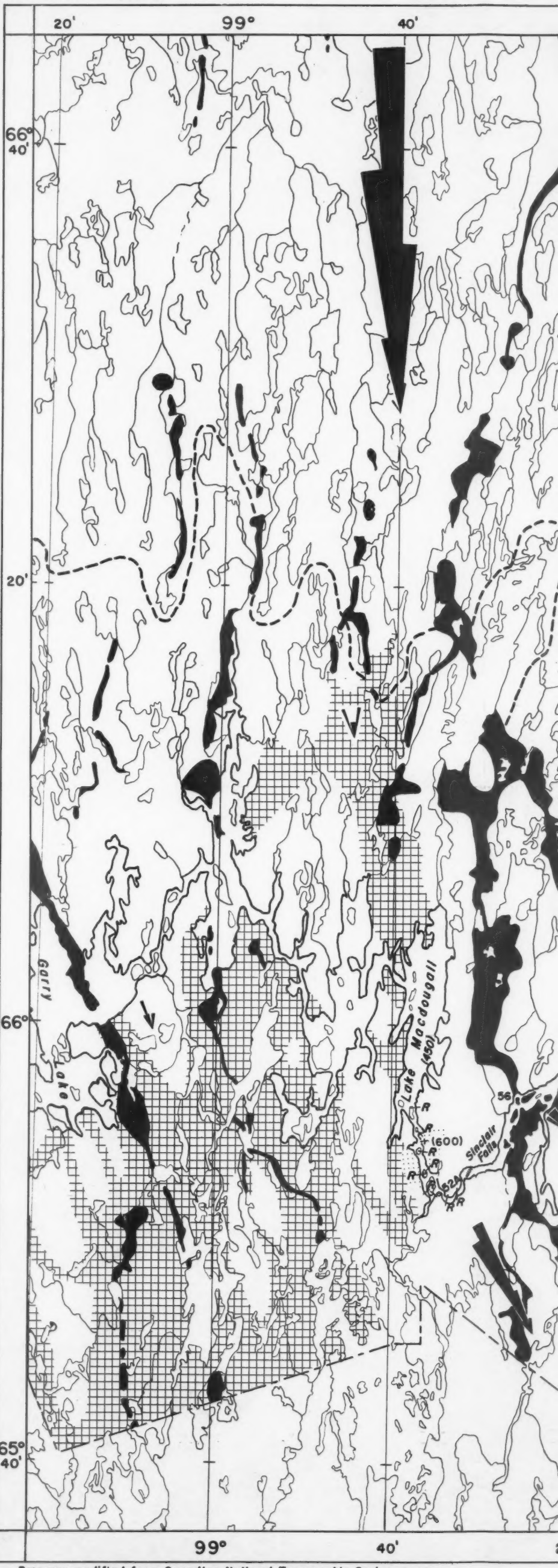

Base map modified from Canadian National Topographic Series maps: Aberdeen Lake N.W. 64/104 and Ogden Bay N.W. 66/104, 1953 editions.

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GLACIA

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BACK

Escape Rapids

Eskimo camp

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Wolf Rapid

RIVER

Eskimo camp

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GLACIAL GEOLOGY OF THE MIDDLE BACK RIVER REG

20' 98° 40' 20'

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MONTRESOR RIVER

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Eskimo camp

Wolf Rapid

RIVER

Back River

Upper Sharp Station

Eskimo camp

20' 98° 40' 20'

GLACIAL GEOLOGY OF THE MIDDLE BACK RIVER REGION

97° 40' 20'

GEOLOGY of the MIDDLE BACK RIVER REGION

District of Keewatin, N.W.T., Canada

66° 40'

SKER DELTAS

ATERIALS
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< 1/3

1/3-2/3

≥2/3

OMALOUS AREAS, origins uncertain

STATION location and number 19 •

ROSE DIAGRAMS of FEATURE ORIENTATION

5° equals one groove, roche moutonnée, drumlin, or one observation on a striae set

Drumlins

Striae, Grooves, Roches moutonnées

DRAINAGE DIVIDES

LIMITS of useful photography

RAPIDS and FALLS observed in field

ELEVATIONS, approximate
(570) Land; *(100) Water*

0 4 8 12 16 miles
Scale

Datum is mean sea level

20'

(595)

98

14B

14A

BASE CAMP

15

(570)

Mt Meadowbank

(535)

14

12

R

(1395)

(100)

Eskimo camp

047

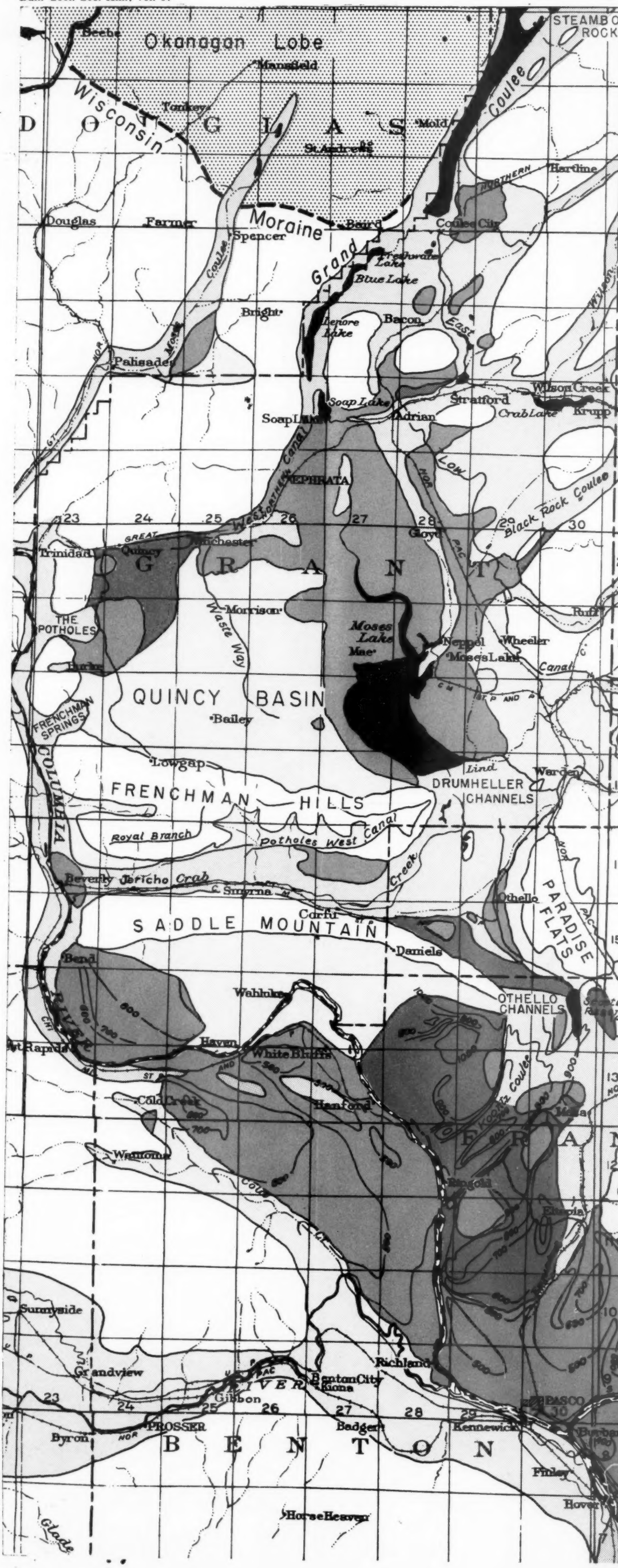
66°

MEADOWBANK

RIVER

65° 40'

97° 40' 20'


Okanogan Lobe
Beebe
Mansfield
Wisconsin
Tonkey
Mold
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St Andrews
Coulee
Hartline
Douglas
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Moraine
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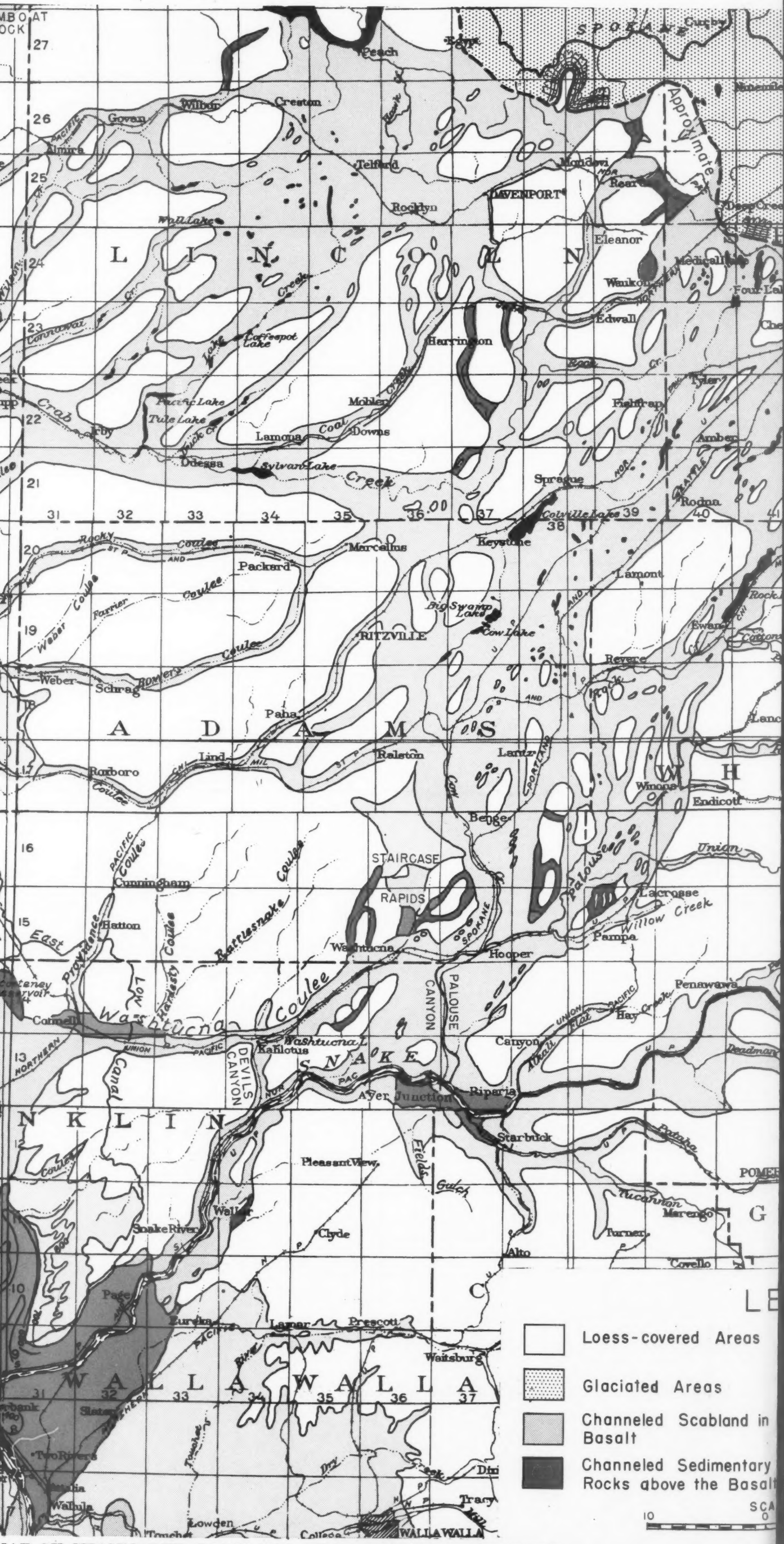
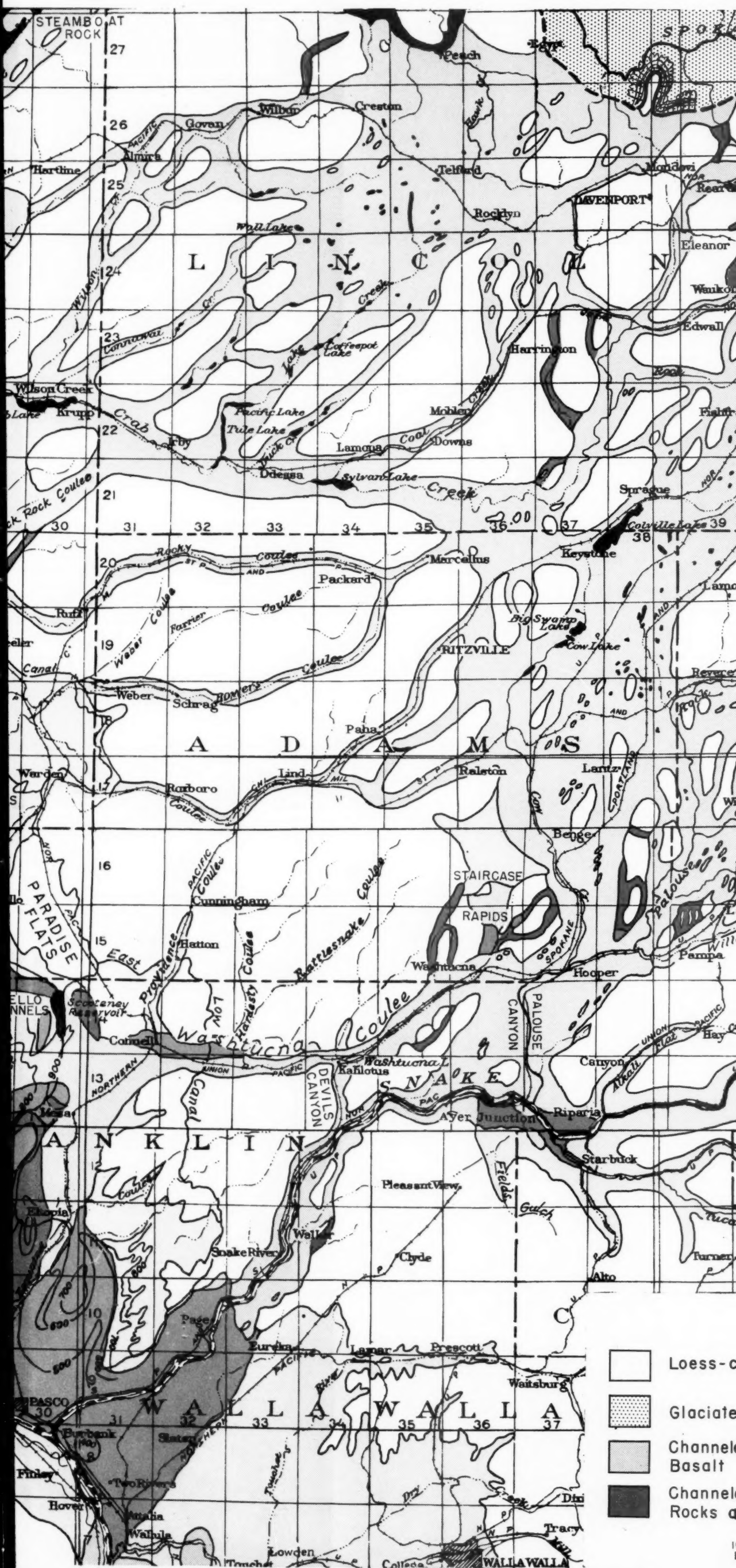
Spencer
Coulee City
Grand
Blue Lake
Bright
Bacon
Palisades
Soap Lake
Adrian
Stratford
Wilson Creek
Crab Lake
Krupp
SOAP LAKE
EPHRATA
Low
Black Rock Coulee
Trinidad
Quincy
GRANT
Morrison
THE POTHOLES
Waste Way
Moses Lake
Neppel
Wheeler
Moses Lake
Ruff
QUINCY BASIN
Bailey
FRENCHMAN SPRINGS
Lowgap
Lind
Warden
DRUMHELLER CHANNELS
FRENCHMAN HILLS
Royal Branch
Potholes West Canal
Beverly
Jericho
Crab Creek
Smyrna
Othello
PARADISE FLATS
Corfu
SADDLE MOUNTAIN
Daniels
Bend
Wahluke
OTHELLO CHANNELS
Rapids
Haven
White Bluffs
Hanford
Wanawa
Royal
Ephrata
Sunnyside
Grandview
Richland
Benton City
Kiona
Gibbon
PASCO
PROSSER
Byron
Badger
Kennewick
BENTON
Finley
Horse Heaven
Glade

MAP



MAP OF CHANNELED SCABLAND OF EASTERN WASHINGTON



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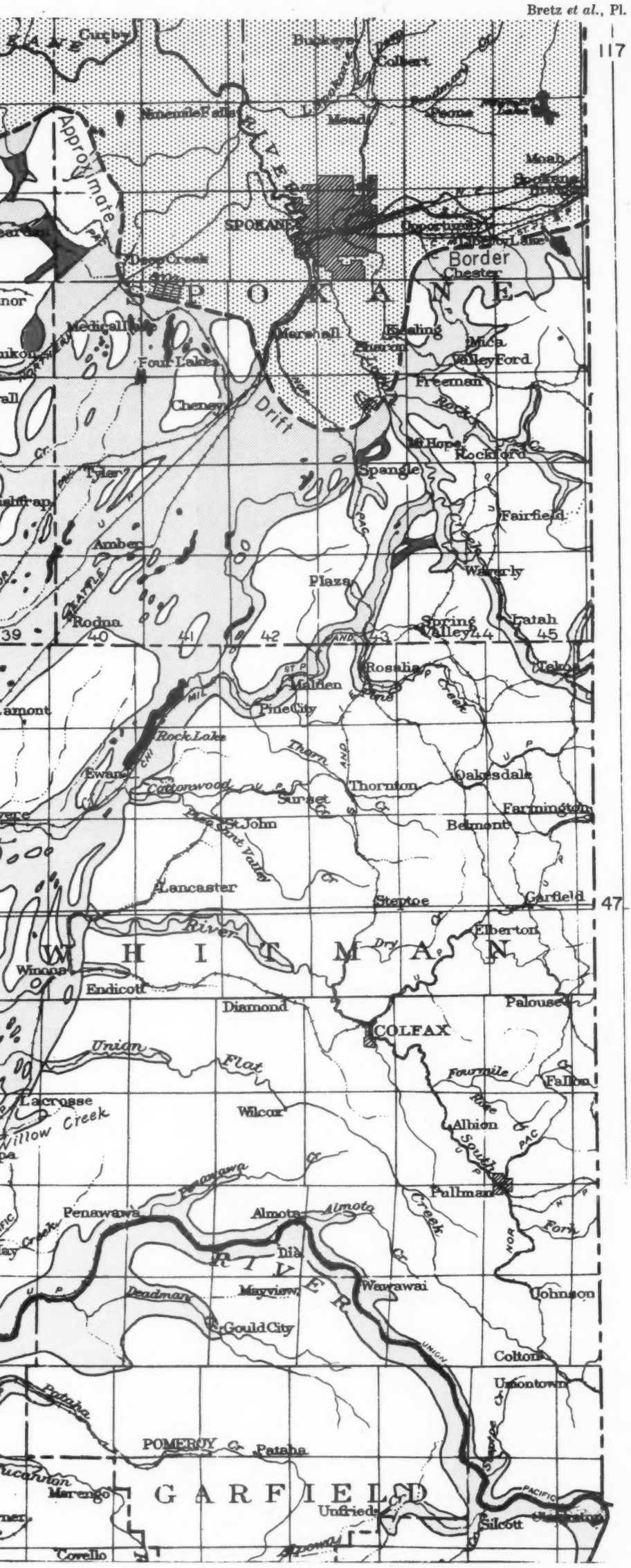

LEGEND

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Glacial Gravel Deposits

Glacial Silt and Sand Deposits

Lakes and Reservoirs

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BULLETIN OF THE GEOLOGICAL SOCIETY OF AMERICA
VOL. 67, PP. 957-1049, 25 FIGS., 16 PLS. AUGUST 1956

CHANNELED SCABLAND OF WASHINGTON: NEW DATA AND INTERPRETATIONS

BY J HARLEN BRETZ, H. T. U. SMITH, AND GEORGE E. NEFF

ABSTRACT

The existence of four different interpretations of that extraordinary assemblage of erosional and depositional land forms of eastern Washington, the "Channeled Scabland", indicates that rigorously definitive diagnostic characters had not been found. This study, dealing with new data, largely from extensive excavations and detailed topographic maps made by the U. S. Bureau of Reclamation in developing the great Columbia Basin Irrigation Project, returns to the earliest of the four interpretations: that channeled scabland is almost wholly the consequence of catastrophic flooding of glacial water across this part of the Columbia Plateau which remade preglacial valleys into an anastomosing complex of great river channels with huge cataracts, deep rock basins, and bars attaining magnitudes unknown elsewhere on earth.

The new evidence is held to establish firmly the following points:

(1) Some structural basins of the region did not have exterior drainage prior to arrival of glacial water.

(2) The gravel hills called bars by Bretz (1928a) have the shapes, surface markings, structures, and topographic situations possible only for subfluvial constructional deposits. In magnitudes and bouldery composition, they are *sui generis*.

(3) Several episodes of catastrophic discharge have occurred across this part of the Columbia Plateau.

(4) The Columbia Valley skirting the plateau has had comparable floods in which the scabland complex did not share.

(5) Successive floods have been differentiated only by topographic relations of their records, not by differential weathering and erosion.

(6) Bretz did not overestimate the magnitude of the erosion by glacial waters.

(7) The existing scabland features contradict the three later interpretations.

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Introduction

The plateau character of eastern Washington is determined by the Miocene Columbia River basalt flows. In northern and eastern marginal portions, islandlike hills of much older crystalline rocks project above the general plane surface. Locally, folding and faulting of the basalt have made strongly expressed corrugations, uplifted little-eroded linear tracts, and depressed areas of only incomplete filling. All relief features, whether diastrophic, erosional or depositional, are interruptions of a regional southwestward slope of the plateau basalt flows.

The original low-lying basalt plain received fluviatile and lacustrine sediments—Ellensburg, Latah, early Ringold (?)—before recorded deformation or erosion. After folding and some weathering and erosion, similarly derived sediments were deposited in structural basins and are still flat-lying. The Ringold formation in the Pasco basin is the outstanding record of this time. Considered to be of Pleistocene age solely from its scanty vertebrate remains, it shows no evidence of contemporaneous glaciation.

Overlying all these older formations is an extensive, thick silt mantle, the "Palouse soil." Most of it is clearly of loessial origin, but a detailed history has not yet been deciphered. A rolling to hilly topography, made mostly by stream erosion but some by wind, appears to have existed on this mantle during the time of accumulation of the dust and has subsequently become much more marked. Calcified and reddened zones, even caliche layers, are known deep in the loess. A widespread sub-

surface caliche was formed after loessial deposition had passed its maximum. It exists also on older rocks where not loess-covered.

The scablands are glacial-river courses where the Palouse and other sediments were swept off the basalt chiefly along preglacial drainage lines, and the basalt itself was notably eroded in places. In these glacial-river courses are extensive stream gravels of the same age and origin. Only the northern edge of the plateau was invaded by the Cordilleran ice sheet.

In the lower reaches of Snake River canyon, in Pasco basin of the Columbia, and in lower Yakima and Walla Walla valleys are widespread lacustrine silts—the Touchet beds—also contemporaneous with glaciation in highlands to the north, east, and west. The Touchet silts record a "Lake Lewis" backwater ponding either from a great blockade in the Columbia gorge across the Cascade Range or from a low stand of this part of the continent and an estuary from the Pacific. "Lake Lewis" is only a general term for possibly several Pleistocene pondings of the Columbia and its tributaries in the southwestern part of the plateau.

The term channeled scabland was first used by Bretz (1923a; 1923b) for the remarkably anastomosing system of glacial-river courses distributed over an area approximately 125 by 125 miles on the plateau.[1] Bretz (1927a) noted that:

> "The channeled scablands . . . cover almost 2000 square miles . . . of the plateau. They are elongate tracts, oriented with the gentle dip slope of the underlying basalt flows, mostly bare rock or with a thin cover of coarse basaltic rubble, commonly with canyons in them, and are bounded by steep slopes of the deep loessial soil of the plateau. They constitute a curious anastomosing pattern, the down-dip convergences inherited from an earlier normal drainage pattern and the divergences, equally numerous, produced by crossing of divides of this older pattern. There are hundreds of tracts of the higher loess-covered areas in the scablands, from a fraction of a square mile to many townships in extent, all discontinuous and bounded by the scabland areas. The steep marginal slopes in loess are in striking contrast to the gentle slopes of the older drainage pattern surviving within each isolated tract. Canyons in the scablands are multiple and anastomosing, amazingly so in some tracts, deep canyons and shallow ones uniting and dividing in a labyrinthine fashion about bare rock knobs and buttes unlike any other land surface on the earth. . . . These canyons are interpreted as channels, not valleys, hence the term 'channeled scabland' ".

This definition dealt only with erosional features. But the glacial rivers that "channeled" the scablands also left deposits whose character, the leading theme in this study, is as extraordinary as that of the erosional forms.

The glacial water that made the scabland crossed the preglacial divide along the high northern margin of the tilted plateau, found gradients of 20–25 feet per mile in the preglacial drainage ways leading southwestward down the dip slope, greatly enlarged them, and crossed minor divides among them to make the anastomosis. At some of these crossings canyons 400–900 feet deep were eroded by cataract recession. Empty rock basins more than 100 feet deep and not of plunge-pool origin occur in many places. The glacial rivers which swept the loessial cover[2] off the underlying basalt and left the sharply expressed marginal scarps, descending from the mature topography of the loess "islands" to the essentially bare basalt of the channel floors, are also recorded by great mounds of basaltic debris, chiefly in the Quincy structural basin and in portions of the canyons of Snake and Columbia rivers which discharged westward. Any satisfactory theory for the genesis of this unique assemblage of land forms must take account of all these features and relations.

Four distinctly different explanations, based on field studies, have been proposed during the past 30 years. At least three theories utilize enough apparently supporting field evidence to have been accepted among geologists acquainted with portions of the region, but each has lacked sufficient evidence to eliminate its rivals.

Field study on which this paper is based is the execution of a long-cherished plan by Bretz to restudy the region with the later

[1] After the appearance of 11 more titles by this author and 5 papers by writers who did not accept his interpretation of the origin of channeled scabland, the details of this definition still satisfactorily encompass all erosional and most depositional features of channeled scabland, except for one writer's explanation which has been vigorously rejected by everyone else acquainted with the region.

[2] Where used without qualification in this paper "loess" includes both the thin, fairly recent loess mantling the mature slopes of the "Palouse topography" and the Pleistocene Palouse soil, itself largely loessial in origin.

rival interpretations in mind. The great Columbia Basin Irrigation Project in the western part of the scabland complex has made new data available from the extensive new excavations and detailed topographic maps by the U. S. Bureau of Reclamation. Several new quadrangle topographic maps by the U. S. Geological Survey have appeared, and the entire region has been photographed by the U. S. Dept. of Agriculture. The field work was carried out with three men working together as a check on a "ruling-theory" tendency.

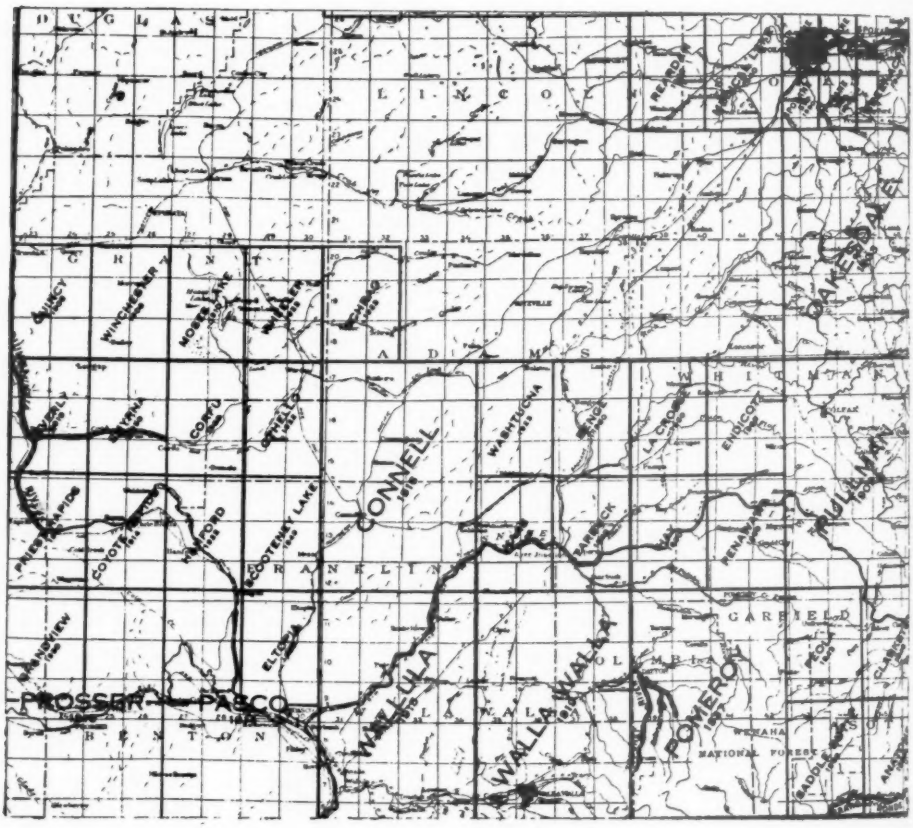

FIGURE 1.—INDEX MAP FOR MOST OF THE QUADRANGLES REFERRED TO IN THIS PAPER
United States Geological Survey.

The finally accepted theory for the origin of this assemblage must consider the entire area. The elephantine character of channeled scabland demands more than cursory inspection, more than detailed study of limited portions. One major phase of the record never before adequately treated is the topography, composition, and structure of the scabland river deposits, the leading theme of the present paper. A. C. Waters, well acquainted with the scabland, wrote the authors that "the best thing you can do to convince your reader is to beg, cajole, even browbeat him into *really looking* at the region's topographic maps". J. H. Mackin, also well conversant with the field, said that "to understand the scabland, one must throw away textbook treatments of river work". A doubting but open-minded

reader is hereby warned. Maps are repeatedly listed for his critical examination.

Acknowledgments

The field expenses of Bretz and Smith have been largely defrayed by a grant from the Penrose Bequest of The Geological Society of America. Additional funds were provided by the University of Chicago and facilities by the University of Kansas. Thirty-four field days in 1952 were devoted by these two authors to this study. Neff, a resident geologist of the Bureau of Reclamation for 5 years, already possessed much critical information and many original ideas, and both the Bureau of Reclamation and The Geological Society of America cordially acquiesced in making him a coauthor after the project grant had been approved. The District Manager of the Bureau of Reclamation, Mr. H. A. Parker, and District Geologist, Mr. W. E. Walcott, generously provided many facilities and granted the use of any geological data the Bureau possessed. I. S. Allison, R. F. Flint, Leland Horberg, Laurence Nobles, A. C. Trowbridge, A. C. Waters, and A. O. Woodford have critically read an early draft of this paper. For all assistance and co-operation, the writers are very grateful. Smith accepted the role in the field of skeptic for all identifications and interpretations. Bretz admits that he served well. However, for the statements in this paper, unless qualified, there is unanimity of opinion among the three authors.

Previous Investigations

(Bretz)

General Discussion

Leighton (1919, p. 34) first recognized the scablands as glacial-river courses. Pardee (1922, p. 686–687) thought the scablands a product of glacial ice, identifying the bouldery, unsorted deposits as till and glacial drift and rejecting floating ice and running water as agents. Bretz (1923–1932) interpreted both the erosional features and the deposits as the result of a brief but enormous discharge of glacial water, termed the Spokane flood. The streamways were interpreted as channels, not valleys. The mounded gravel deposits were called river bars, although far larger than any other known river-channel deposits. Bretz stressed contemporaneity of discharge through all the channels. Jenkins (1925) approved of the concept that the narrow Wallula Gateway, where the Columbia crosses the Horse Heaven Hills anticline below the plateau scabland, had served as a bottleneck for an excessive quantity of glacial water, the Spokane flood, that caused flooding back up Snake River to Lewiston, Idaho, and even farther.

During the nine years when Bretz's papers were appearing, many voices were raised in expostulation, some recorded in Proceedings of the Geological Society of the Washington Academy of Sciences (V. 17, no. 8, 1927, p. 200–211) and in the Journal of Geology (V. 35, no. 5, 1927, p. 461–468). The general tenor was that this heresy must be gently but firmly stamped out.

In the well-organized "discussion" following a lecture by Bretz before the Geological Society of Washington in 1927, W. C. Alden suggested (not printed in the Proceedings above noted) that the rock basins might be collapsed lava caves, and said "it seems to me impossible that ... the Cordilleran ice ... could, under any probable conditions, have yielded so much water in so short a time". He asked for "repeated floodings of much smaller volume" and suggested that contemporaneous development of all parts was an erroneous interpretation. James Gilluly said "that the actual floods involved at any given time were of the order of magnitude of the present Columbia's or at most a few times as large seems by no means excluded by any evidence as yet presented."[3] G. R. Mansfield said "it does not seem to me to be necessary to assume that all the scabland channels were occupied by water at the same time.... The scablands seem to me better explained as the effects of persistent ponding and overflow of marginal glacial water which changed their position or their places of outlet from time to time through a somewhat protracted period." To escape the idea of contemporaneity, O. E. Meinzer suggested that "tilting and folding have occurred ... since the cutting of the

[3] The first five of Bretz's papers were then in print.

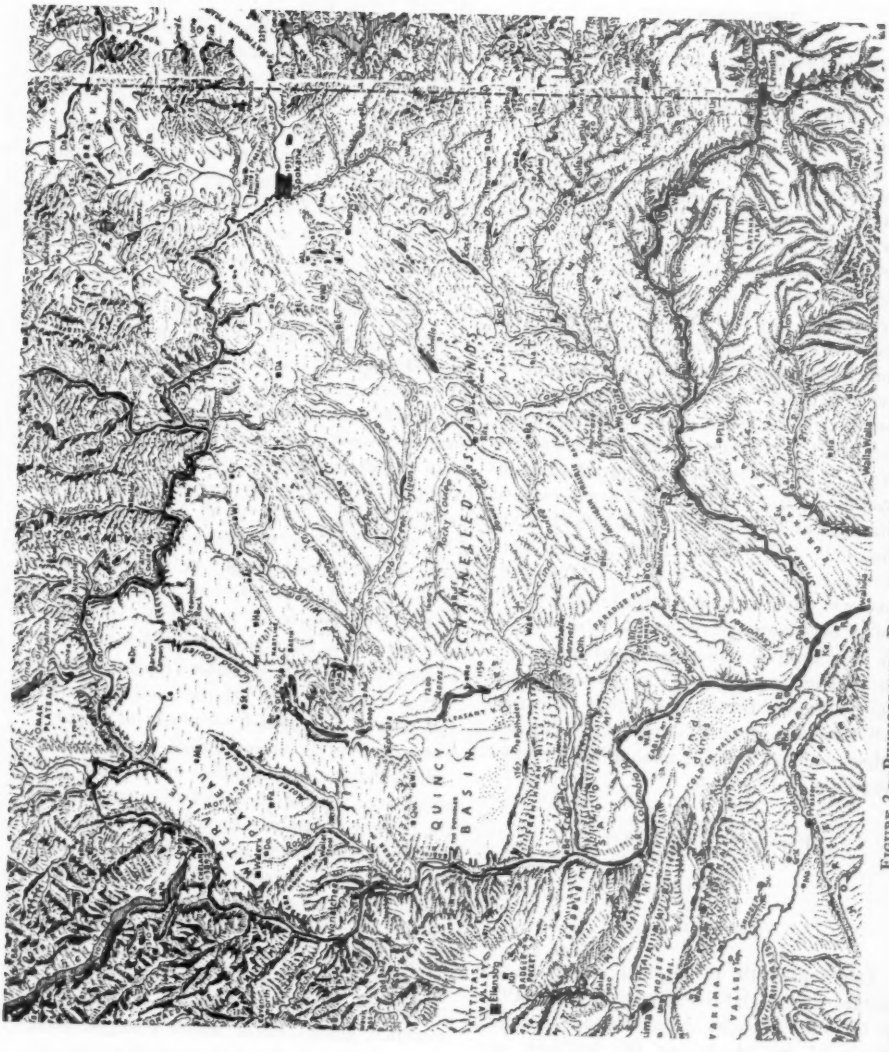

FIGURE 2.—PHYSIOGRAPHIC DIAGRAM OF CHANNELED SCABLANDS AND ENVIRONS
From Ralisz (1941)

Pleistocene channels" (bringing channel heads to the same level) and said that "before a theory that requires a seemingly impossible quantity of water is accepted, every effort should be made to account for the existing features without employing so violent an assumption." G. O. Smith said (not printed in the discussion) that the loessial scarps margining scabland tracts might be the product of rainwash and erosion by wind rising up over lower basalt cliffs. Not printed in the discussion also was a suggestion from H. G. Ferguson that river ice jams may have caused some of the high-level divide crossings in the anastomosis.

E. T. McKnight, who shared in the discussion, later that year (1927) published elaborated objections to the flood theory, he and his sponsors preferring to believe that certain glacial stream courses entering Columbia River in western Franklin County (the Koontz Coulee group) (Fig. 17; Pl. 1) are hanging because that river has widened its valley in postglacial time, not because the glacial Columbia's surface was at or near the level of the coulee mouths when they were made. Like Meinzer's criticisms, McKnight's had the merit of being based on personal experience; he, as an undergraduate, having "held the rod" during the topographic mapping of three 15-minute quadrangles on the plateau.

Bretz responded to these suggestions and criticisms in the printed "discussion" already noted, also in a reply to McKnight (1927a) and in a paper (1928c) devoted to all alternative hypotheses ever suggested at that time. Although several contributed ideas were acceptable modifications of the original Spokane Flood interpretation, a Pleistocene catastrophic flooding was insisted on. Until 1930, however, no plausible source for the enormous quantity of water demanded in so short a time had been found. Bretz later (1930a, p. 92) ascribed the scabland flooding to an abrupt failure of the ice dam which ponded Glacial Lake Missoula (Pardee, 1910, p. 376–386).

Harding (1929), after spending a day with Bretz during the field season of 1928 and learning that Lake Missoula was to be proposed as the source of the flood water, rushed that statement into print, without consultation or acknowledgment, a few months before Bretz's announcement. Allison (1933) proposed that the flooding was caused by a blockade of floating ice in the Columbia gorge through the Cascade Mountains, and that this blockade gradually grew upstream until the normal meltwater discharge from the Cordilleran ice sheet was "diverted by the ice of accessory blockades into a succession of routes across secondary drainage divides at increasing altitudes, producing scablands and perched gravel deposits along the diversion routes. . . ." Allison favored a landslide as the initiating cause of the Columbia blockade, "with freezing of the river (or the water-ice mixture) and the confinement of the ice in the narrow and somewhat crooked gorge as contributing causes." "The formation within the Columbia River gorge of an ice jam, capable of withstanding the pressure of at least 900, and perhaps 1100, feet of water is a herculean requirement but the field evidence seems inescapably to demand it." Field study for his theory was limited to areas south and west of the main scabland.

E. T. Hodge (1934) published a brief paper on the origin of the Washington scablands, listing a long and complicated series of late Tertiary and Pleistocene events in the region, involving ice jams during three glaciations, but did not support his proposed history with field evidence.

Flint (1938), studying especially the Cheney-Palouse tract, denied that glacial-stream water attained any great volume at any time or place, asserting that "this discharge down the Cheney-Palouse tract was less than that of the Snake River today." His thesis was that during normal discharge of Cordilleran meltwater[4]

[4] Flint published (1937) a revision of the drift border as shown in Plate 1 of this paper; he placed it farther south so that it overlapped scabland at the head of the Cheney-Palouse tract. This revision was based on two criteria; (1) smoother slopes and lower relief of the northernmost Palouse hills, and (2) glacially derived erratic boulders and patches of till in the added territory. However, typical steeply sloped, fine-textured Palouse topography in places far from the glaciated area grades into tracts of lower relief and gentler slopes. Moreover U. S. Production and Marketing Administration's mosaics 1, 5, and 6 for Lincoln County, Washington, show no contrast in topography or land use between the two sides of Flint's minutely lobate glacial margin across this tract. Instead, a marked belt of elongated, subparallel loessial hills south of this boundary extends with equal development, spacing,

FIGURE 2.—PHYSIOGRAPHIC DIAGRAM OF CHANNELED SCABLANDS AND ENVIRONS
From Raisz (1941)

rising ponded water in Pasco basin (Lake Lewis) at the confluence of the Yakima, Snake, and Columbia rivers (Fig. 17; Pl. 1) caused a gradual aggradation with silt, sand, and gravel over the Cheney-Palouse tract and brought about the divide crossings. Implicit in his view, although not specified, were comparable deposition and interfluve crossing throughout the entire scabland. Lowering of Lake Lewis then allowed the continuing normal discharge to remove the fill in large part, to scarp the bordering loessial hills, and to make the scabland. Various specific aspects of orogeny, epeirogeny, volcanism, and gradation were explored as causes for the ponding, and "valid possibilities" were seen in a dam made by landslides or by a Mt. Hood valley glacier.

Allison (1941) found Flint's fill hypothesis out of harmony with the constructional forms and stratification of the mounded gravel deposits, incorrect in correlating Lake Lewis with deposition of most of the gravel deposits in Snake River canyon, and demanding of the streams eroding the fill "decidedly abnormal, if not impossible, behavior" along the Cheney-Palouse tract. He again urged that only the ice-jam theory could explain the channeled scabland.

Pardee (1942) found high-eddy deposits, bar-like forms, giant current ripples, strongly marked channels, and severely scoured salients in some of the valleys occupied by Glacial Lake Missoula in western Montana. He postulated failure of the lake's ice dam, which caused a sudden, great lowering. Strong currents are indicated by unique giant current-ripple marks that attain amplitudes of 50 feet and have wave lengths of as much as 500 feet. Pardee said nothing in this paper about disposition of the 500 cubic miles of water the lake contained. However, it escaped westward around the north end of the Bitterroot Range into Idaho and thence could have gone only down the Spokane Valley toward the heads of the channeled scabland.

Hobbs (1943; 1947) presented another theory, stating that (1) a lobe of the Cordilleran ice sheet had almost covered the region, (2) scabland was a product of glacial scouring, (3) the rock basins were glacially plucked, (4) the loess, generally considered prescabland in age, was actually deposited by anticyclonic winds off the ice during retreat of his lobe, (5) mutilated moraines recorded at least five stages of retreat, and (6) the channel complex was the product of glacial-border drainage, each retreatal stage developing new channelways across freshly exposed scabland. Hobbs' interpretation of 1943 (unaltered in 1947) was based on inadequate field study and is considered to be invalid. The Palouse loess occurs in hundreds of scarp-bounded, island-like tracts separating or interrupting the scabland channels, and these scarps sharply truncate the older mature topography of such islands, leaving many of their valleys hanging. Reddened zones and caliche layers with root casts and molds are buried deep within the loess. Hobbs was the only investigator who interpreted any land forms of the scablands as moraine relics, and he made only the bland statement that they exist in specified places, citing no evidence to show that they are glacially built deposits and making no comment on how earlier students could have missed them. He actually described a "moraine" in the bottom of one of the channels (Ephrata) for Grand Coulee water across the Quincy gravel fill. Regarding the sequence of border drainages, he required his glacial rivers successively to abandon open *lower* routes for *higher* ones as his lobe retreated, and he used one route (Providence Coulee) that has no scabland, glacial river gravel, or erratics and is far too high for his purpose. Finally he left some of the best-developed scabland (Drumheller, Othello, Moses Coulee, Lower Crab Creek, Palouse-Snake divide, Columbia and

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and orientation northward for 12 miles into his glaciated area. Inspection of the northwestern portion of the Rearden quadrangle map (U. S. Geol. Survey, 1942) will bear out this criticism. Also many glaciated erratic boulders have been found along all scabland river routes, along the Columbia beyond the scabland, and, indeed, back-floated up the Willamette valley of western Oregon. Furthermore, large berg "nests" of these boulders with an accompanying matrix of till are especially numerous in Pasco basin 75 miles or more down the Columbia from the southernmost known glacial drift *in situ*. Because of these facts, Flint's revision does not carry conviction, and the 1928 mapping of the drift border by Bretz is followed in Plate 1. In the absence of any end moraine and of well-marked ground moraine on the plateau east of Grand Coulee, precision in mapping the drift border for this area probably is impossible.

Snake River valleys) outside the limits of his lobe.

R. L. Lupher (1944), dealing with the Touchet beds—*i.e.*, fine sediments in contiguous parts of Snake, Walla Walla, Yakima, and Columbia valleys, ascribed by Flint and Allison to deposition in Lake Lewis—briefly discussed the origin of Washington scabland. He found objections to the interpretations of Bretz and Flint and accepted Allison's ice-jam theory.

A reader familiar with Bretz's earlier publications will find several revisions of interpretation in this paper, the most striking one being abandonment of the term "the Spokane Flood" and a demand by all three authors for several scabland floods. Obviously the present paper cannot be simply a series of assertions and denials. Field evidence must be presented in sufficient detail to allow the reader to judge of its appropriateness and adequacy for accompanying interpretations.

Summary of Original Interpretations

(1) The Palouse loess of the basalt plateau is older than the scabland. (Accepted by all critics and writers except Hobbs, who thought it essentially contemporaneous.)

(2) The scabland channel complex has been made by glacial water entering and modifying preglacial drainage lines on the loess-covered plateau. (Only Hobbs has a variant interpretation.)

(3) The scabland rock basins have been made by glaciofluvial erosion, chiefly by plucking in the columnar basalt. (Hobbs has differed, asserting glacial plucking; Alden has suggested that they might be collapsed lava caverns.)

(4) The downstream-facing cliffs interrupting channel floors are abandoned cataracts. (Denied by no critic.)

(5) The loessial scarps bordering scabland channels are marginal steepenings made by glacial rivers. (Only G. O. Smith has expressed doubt.)

(6) The scabland gravel deposits are contemporaneous with the channeling. (Flint demurred, believing the channels were made during removal of former complete fluvioglacial valley fills. The critical items here concern the ability of Flint's low-gradient streams of moderate size to detach great boulders from bedrock, to transport them for miles, then to remove the fill and to make the rugged, channel-bottom scabland topography associated with "residual" hills of gravel. Without a contemporaneous "Lake Lewis", whose rise caused the aggradation and whose immediate subsidence caused the erosion, Flint's theory has no logical foundation. Demonstration that the gravel "remnants" are stream-cut terraces is fundamental for his theory, and scabland forms buried under gravel are unallowable.)

Allison postulated a prescabland gravel-terrace system in Lower Snake River canyon, erosional remnants of which were subfluvially shaped by the glacial water from the scablands. (If correct, there must be comparable but unmodified terraces in the canyon upstream from entrance of the scabland rivers. In the terraces modified by the glacial Snake, there should be abundant well-worn Snake River-type pebbles. In unmodified cores of these deposits cementation and/or weathering would be accessory evidence.)

(7) The structure, composition, and topographic expression of these deposits record subfluvial origin as river bars of gigantic size. The undrained fosse depressions among and behind these gravel hills, the midchannel location of many, and the numerous huge current-ripple marks conforming to their shapes are undeniable evidence for bar origin. (Denied by Flint, Gilluly, Hodge, and the sponsors of McKnight. Flint claimed that they are terraced remnants of once-continuous fills, and Gilluly called their long foresets evidence for deltaic origin; Allison accepted them as subfluvial forms, although his concept cannot explain the mid-valley location of many.)

(8) The similar altitude of the upper limit of glacial water in all scabland channel heads and the anastomosing pattern indicate approximately contemporaneous functioning at some time over the entire area. (Strongly questioned by Alden, Gilluly, Hodge, Mansfield and Meinzer, accepted by Flint and Allison. Evidence for contemporaneity at some stage in scabland history must be expected at convergences of channels with separate heads.)

(9) The many minor divide crossings were due to a sufficiently great volume of glacial water to fill pre-existing valleys up to overflow sites. Earlier valley-train deposits may have been a factor in overflow. (Denied by Alden, Flint, Gilluly, Hodge, Mansfield, and Meinzer. Flint specifically shallowed the preglacial valleys with fluvioglacial deposits caused by deep backwater (Lake Lewis) beyond the plateau until spillover into adjacent valleys occurred; this before basalt scabland was made. Allison used local ice jams to make local impoundings and spillovers. There appears to be no third reasonable alternative. Evidence for the flood concept must include demonstration that a general fill never existed and that ice jams left no definite local records of their actual existence.)

(10) Ice-rafted stones and water-transported finer sediments back in valleys, tributary to scabland channels, that did not carry glacial-water discharge were the consequence of back-flooding because of the great volume and depth in the main channels. (Denied or ignored by all but Allison and Flint.)

(11) Bars, high-level channels, tributary valley-mouth deposits, and a "Portland delta" record the flood waters descending the Columbia beyond the plateau, crossing the Cascade Range and entering a Columbia valley estuary. (Allison, 1933, and Hodge, 1934, denied the Portland deposit as a delta; Piper, 1932, denied the flood origin of tributary mouth deposits near The Dalles, Allison and Flint believed a dam (landslide? Mt. Hood glacier? ice jam?) blocked the Columbia Gorge across the Cascades. High-level, berg-rafted, glaciated boulders have been floated to position in lower Columbia and Willamette valleys but perhaps are not all contemporaneous with scabland making. The dam must remain hypothetical until definitive evidence for its location is found on the walls of the gorge.)

(12) Only a great flood could have caused simultaneous discharge of the three great cataracts and one enormous high-gradient spillway out of Quincy basin. (Simultaneity denied or ignored by all opponents of the flood hypothesis, and the four accordant initial discharge levels explained by Meinzer as a fortuitous leveling up by subsequent earth movements. The only other alternatives are that ice jams blocked three of the four dischargeways at a time, in a rotating sequence, or that the basin once had an alluvial fill on which the glacial stream shifted from one dischargeway to another.)

(13) Upper Grand Coulee's dry canyon was made by cataract recession completely through a preglacial divide. (Denied by no critic.)

(14) Water for scabland making came from the bursting of the glacial dam impounding Lake Missoula in western Montana. (Rejected by all except Pardee who discovered evidence for torrential emptying of a large part of the lake. There is little field evidence at or near the dam site to prove a catastrophic bursting, but the lake water certainly discharged westward. Acceptance of evidence for scabland floods entails acceptance of the strong probability of a direct cause-and-effect relationship. (Since this was written, Alden has indicated a "perhaps" acceptance. See footnote 43.)

(15) The scablands east of Grand Coulee were ascribed to a pre-Wisconsin (Spokane) glaciation and Grand Coulee's completion to flood discharge during the Wisconsin glaciation. (Hodge announced three different glaciations during scabland development, while Flint claimed that the Spokane drift is Wisconsin in age.)

Advance Summary of New Evidence and Interpretations

Changes and elaborations in Bretz's earlier interpretations include:

(1) tentative identification of some gravels, both high-level and buried, as remnants of prescabland normal outwash;

(2) recognition of a heavy caliche on the loess and older deposits in part of the region as the marker for immediately prescabland topography;

(3) discovery of giant current-ripple marks on bar deposits in nine different channel-ways, distributed in more than a dozen localities ranging from the northern to the southern, and from the eastern to the western sides of the complex;

(4) establishment of the river-bar origin of the discontinuous mounded gravel deposits in scabland channels from aerial photographs

taken by the U. S. Agricultural Adjustment Administration, the U. S. Production and Marketing Administration and H. T. U. Smith, from the detailed topographic maps of the U. S. Bureau of Reclamation, and from many extensive new excavations;

(5) recognition of the occurrence of several successive floods, the earliest recorded only in the Columbia valley, the second involving perhaps the entire area, later ones restricted to channels already deepened, and the last one again limited to the Columbia valley;

(6) recognition that Touchet silts were not contemporaneous with scabland making.

Grand Coulee and Quincy Basin

Maps: Grand Coulee (4 sheets), Moses Lake, Othello, Quincy and Winchester quads., U.S.G.S.; R4-5720, sheets 1 and 2, U.S.B.R.

(Bretz, Smith, and Neff)

Grand Coulee

A tandem canyon 50 miles long and more than 900 feet in maximum depth, Grand Coulee is the greatest scabland channel. It is a glacial Columbia River's detour when the river's preglacial course was dammed by a lobe of ice pushing southward from the capacious Okanogan Valley (Pl. 1). Striae and moraine relics near the head of Grand Coulee show that it had present depth before the last invasion by the Okanogan lobe. Grand Coulee has been generally accepted as largely a recessional cataract gorge (Bretz, 1932), its northern half eroded across a previously intact divide, its southern half excavated along the strike of the steep limb of an asymmetrical flexure, the Coulee monocline.

The northern, or Upper, Coulee was initiated where the glacial stream cascaded some 800 feet down the steep southeastern slope of the monocline.[5] When this cascade migrated north of the fractured and steeply tilted zone into high, flat-lying basalt, it became a typical recessional cataract (Steamboat Falls) nearly 900 feet high. Eventually its retreat extended the lengthening gorge across the divide into the preglacial Columbia River valley, and thus the cataract destroyed itself. Scabland records of the glacial river, made before much recession had occurred, exist on the rim of the gorge nearly 1000 feet above the floor. Subsidiary cataracts along both walls, one of them as wide as Niagara and three times as high, tell something of the breadth of the Grand Coulee glacial river before cataract recession narrowed it to a minimum of a mile. (*See* Bretz, 1932.)

These brief items[6] will indicate the magnitude of this scabland river across the Coulee monocline. Along the lower or southern half of Grand Coulee the stream spread widely, enormously enlarged three small, preglacial drainageways, and made a huge new gorge by quarrying out the fractured basalt along about 10 miles of the monocline's steep limb. Minor structures involved in this anastomosing course were four nearly east-west anticlines (High Hill, Trail Lake, Pinto Ridge, and Soap Lake) with separating synclines. Their higher portions stood above the wide-spreading river, but lower parts were crossed by floods and greatly eroded to make five additional canyoned courses (Jasper, Lenore, Unnamed, Dry, and Long Lake) over a width, including the monoclinal course, of about 15 miles. The discharge of all these entered the extensive Quincy structural and topographic basin, largely covered it with basaltic gravel, and made four immense simultaneously operating scabland dischargeways (Crater, Potholes, Frenchman Springs, and Drumheller) leading to the Columbia valley west and southwest of Quincy basin.

Trail Lake Anticline and Bacon Syncline

One of the most significant tracts of this Lower Grand Coulee area is the denuded portion of Trail Lake anticline with Bacon syncline to the south (Pl. 2). Over a width of 5 miles, the glacial floods swept up the gentle northwestern anticlinal slope and descended the steep southeastern slope to enter the pre-

[5] Bureau of Reclamation exploratory drillings found a gravel-filled plunge pool at this place whose bottom is 300 feet lower than the scabland floor immediately downstream, just south of the flexure.

[6] The conclusions regarding Grand Coulee are based on details set forth in another publication (Bretz, 1931). Not having been specifically challenged by critics, the field evidence will not be repeated here.

glacially undrained Bacon syncline.[7] Apparently, this structural basin was still filled with superbasalt sedimentary rock, capped with loess. On a basin floor at about 1650 feet A.T., the first glacial discharge across the anticlinal crest deposited 100 feet of basalt gravel (with some erratics) just south of the crest, burying a loess with a calichified upper zone. The altitude of this high gravel flat proves that there was, as yet, no deep Lower Grand Coulee and no Dry or Long Lake (Spring) Coulee outlet notches in the syncline's rim.

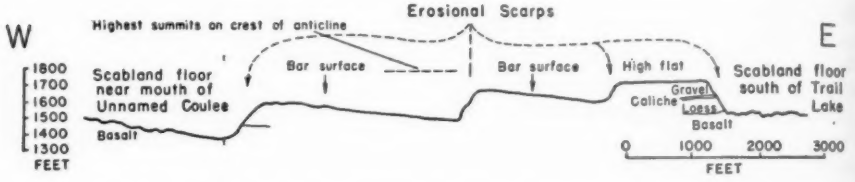


FIGURE 3.—PROFILE OF GRAVEL DEPOSITS ON SOUTH SLOPE OF TRAIL LAKE ANTICLINE

This scabland gravel deposit in Bacon syncline was once called a bar (Bretz, 1932, p. 28). The lower southern and western parts are mounded bars (Fig. 3), but the high flat-topped portion is only a remnant of a larger deposit (Pls. 2, 3), perhaps an outwash plain; its present barlike outline was the result of erosion by later glacial discharge across the Trail Lake anticline when Dry and Long Lake coulees were being notched into the rim of the Bacon syncline. Neff reports small remnants of gravel on the north slope of Pinto Ridge (the southern side of the syncline), at about the same altitude.

Deepening of these notches is considered to have caused eventual removal of most of this gravel and the underlying sedimentary material in the structural basin from the steep southern basaltic limb of the anticline. Three or four cascades began to concentrate the sheet of glacial water, and these, receding north of the axis, became cataracts. One, at the head of the Unnamed Coulee, had retreated about 4000 feet when the last discharge to cross the anticlinal crest subsided. It is a splendid example of an abandoned waterfall with three alcove heads (Pl. 3) whose cliffs are 100–150 feet high, a cataract supplied only by a sheet of water flowing diagonally updip across the unchanneled, gentle northern slope of Trail Lake anticline.[8]

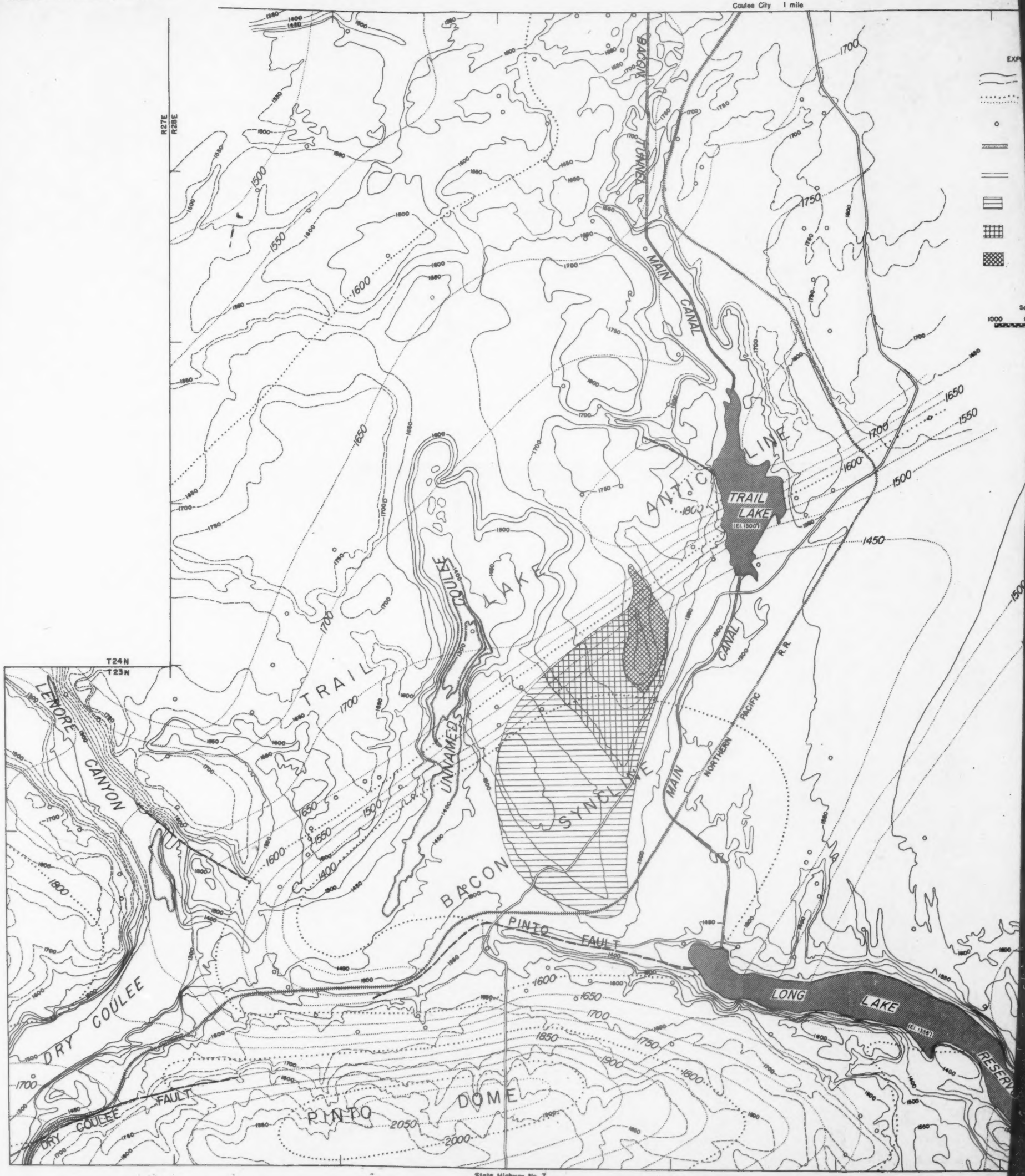
Bacon synclinal basin apparently carries a record of four episodes of glacial-water discharge in distributary fashion out of Upper Grand Coulee across Trail Lake anticline: the first recorded by the high gravel flat, the second and third by the two broad bars just to the southwest which lie 50–200 feet lower, while the fourth completed the Unnamed Coulee cataract's recessional gorge immediately west of the scarp of the lowest bar and 150–200 feet still lower. All three scarps in this gravel-covered tract are erosional, but only the highest can be considered a terrace scarp.

Bacon syncline's bottom scabland at the southwestern outlet hangs between 100 and 150 feet above the floor of Lenore Canyon-Dry Coulee with no suggestion of cataract development at the junction. This seems to record deepening of the canyon-coulee after the last flooding of Bacon syncline. If so, the great Dry Falls cataract group (400 feet high) at the head of Lower Grand Coulee (Bretz, 1932) is younger than the Unnamed Coulee cataract, and a fifth episode of discharge down Grand Coulee is indicated.

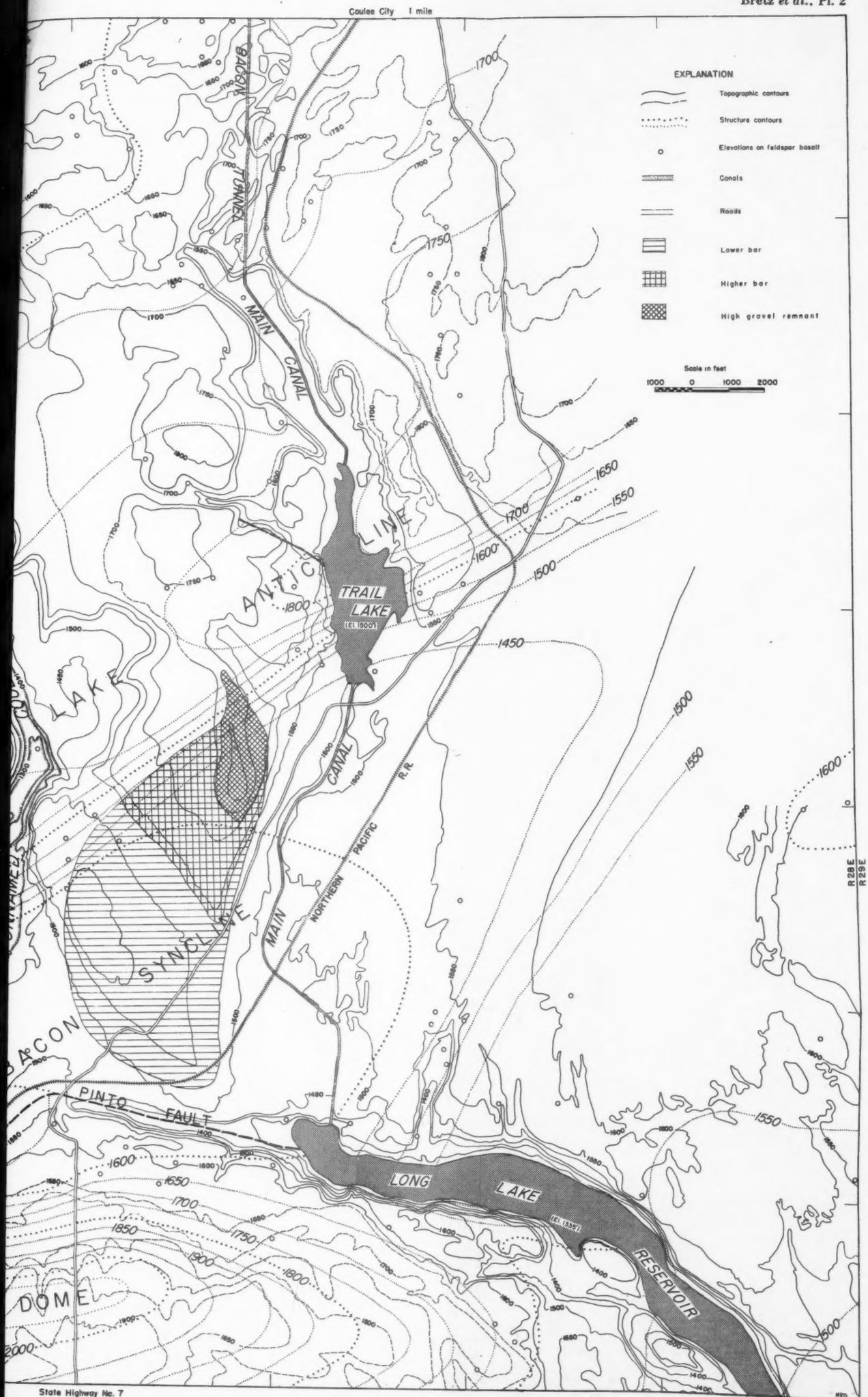
It may be that Lenore Canyon never carried

[7] Discovery of two feldspar-bearing flows in the Columbia River basalt of eastern Washington (W. H. Irwin, unpublished manuscript) has made possible a detailed structural mapping by the Bureau of Reclamation for the Columbia Basin Project. The Bureau's map shows that the upper feldspar flow in Bacon syncline is 1375 feet A.T. in the central part and 1700–1775 in the eastern and western low saddles. Their 2-foot interval contour map shows breaching only by glacial-river channels cut into these saddles.

[8] Another, at the head of Long Lake Coulee, is closely paralleled by the Main Canal. Despite a tunnel about 9000 feet long through the anticlinal crest, the canal water, after emerging, drops 165 feet in half a mile to pass the site of the ancient cataract.



MAP OF TRAIL LAKE ANTICLINE AND VICINITY, WASHINGTON



AKE ANTICLINE AND VICINITY, WASHINGTON

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a through stream from Lower Grand Coulee to Dry Coulee; that floods across the ravaged country north of the canyon were divided by the unsubmerged portion of High Hill, part going west to Lower Grand Coulee, part east to Dry Coulee. Supporting this idea is the narrowness and lack of bars in mid-length of the canyon, and the absence of a continuous gradient.

Soap Lake

The Soap Lake anticline, southernmost subsidiary spur trending eastward from the Coulee monocline, had a small, antecedent drainageway across it which glacial waters obliterated in making the lake basin. However, several miles of this preglacial streamway can be identified farther north, east of the great monoclinal axis (Bretz, 1932). The Bureau of Reclamation found rock bottom 214 feet below the lake surface, with 109 feet of glacial-river gravel above it and 84 feet of postglacial mud above that.

Gravel deposits of the Quincy basin constitute the notched southern rim of the basin 75–250 feet above the present lake surface (Figs. 4, 5). Thus rock bottom in the mouth of Grand Coulee is 290 feet below the lowest gravel surface 2 miles to the south. Were the rim restored, and the fill under the lake removed, there would be a hole in the glacial river channel 340–465 feet deep.[9]

The mouth of Lower Grand Coulee, therefore, has discharged a prodigious stream out on the lower land of Quincy basin (Fig. 4). Because tributary Long Lake and Dry coulees probably once functioned at the same time, and their combined widths near the mouths are about 1½ miles, any estimate of the magnitude of Lower Grand Coulee river at a late functioning must be correspondingly increased.

Grand Coulee water, on emerging from its rock-bound course, spread widely in Quincy basin, and the extensive gravel deposit there records a great decrease in transporting ability. The back slope, northward down to Soap Lake, is a smooth, subfluvial, equilibrium profile lacking in ravines, terraces, scarps, channels, bars, or the hummocky topography left by stranded ice blocks (Waters, 1933). It is a slope adjusted to the volume, velocity, and load of that great river's last flood. Was this slope erosionally developed in previously deposited gravel, or is it a depositional profile of that glacial flood?

Quincy Basin

At least 500 square miles of this large, structural depression are covered with gravel and sand brought in chiefly by Grand Coulee and Upper Crab Creek glacial rivers (Pl. 1). The maximum known thickness is 128 feet. The gravel, like that elsewhere in this Washington complex of glacial streams, is but little weathered and in but few places is incipiently cemented. It is characteristically bluish black, about as "bright" as basalt aggregate from a crusher. This is true for the oldest gravel as well as for the youngest. Except in one place, the amount of weathering since deposition cannot be used to indicate relative age. Freshness of the gravel indicates derivation from below the zone of preglacial weathering, estimated by Neff to average 25 feet in the irrigation-project area. Stained and partially indurated clays and sands containing shells, bones, and wood crop out in places along the basin's margins (Schwennesen and Meinzer, 1918). These are older than the scabland gravel. Above the level reached by glacial streams, near Winchester and Quincy and south to Frenchman Hills, these older sediments carry a caliche as much as 15 feet thick. In many places throughout the scabland a similar caliche overlies loess, Ellensburg, Ringold, Columbia River basalt, and cemented old gravels. Scabland sand and gravel deposits commonly contain fragments of caliche but nowhere possess a caliche cap *in situ*. A heavy caliche is believed to dis-

[9] Silts containing fresh-water shells along Lake Lenore and Soap Lake record a postglacial humid climate when the Soap Lake depression held a much larger lake with overflow down Rocky Ford Channel. Such a lake must have extended northward almost the full length of the Lower Coulee. The closed depression containing Soap Lake thus includes all the other lakes for 25 miles to the north. Present infiltration of irrigation water is raising the level of Soap Lake and, unless checked by increased evaporation from expanded surface and by interception wells, may submerge the city of Soap Lake and reestablish a continuous lake in Lower Grand Coulee.

tinguish the prescabland surface over much of the region.

Although the surface of the Quincy basin's scabland gravel deposit has an east-west cross-sectional relief of nearly 200 feet, owing mostly to later channeling, all parts south of the notched rim of Soap Lake basin have a fairly uniform, gradual, southeasterly slope down toward Drumheller Channels, the largest and deepest of the four glacial spillways out of Quincy basin. The basin fill has no slope toward the three western cataracts.

Three definite channels traverse this southeasterly slope of the gravel plain, the lowest of which, Rocky Ford, contains Moses Lake in its southern part. Its deepest part is sinuous like a river channel, but its broad higher floor is a complex of anastomosing smaller channels.

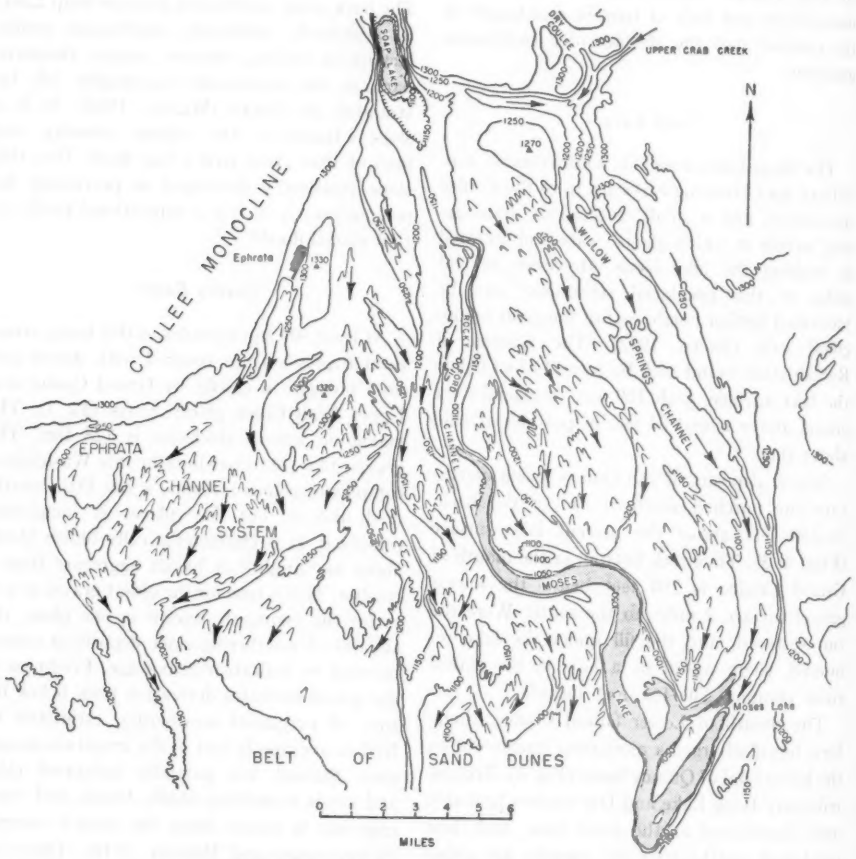


FIGURE 4.—MAIN PORTION OF QUINCY BASIN

Showing the three channels leading from Soap Lake across the gravel fill and the Λ-shaped heads of minor channels. From U. S. Bureau of Reclamation topographic map R4-5720, sheet 1. Contour interval 50 feet.

A barchan-dune belt has dammed the deeper central part of this channel and made the lake. This ponding has brought out a peculiarity possessed also by Ephrata channel, the highest of the three—the existence of tributary channels (Parker, Lewis, and Pelican horns or bays) which head on the gravel plain south of the rim of Soap Lake basin. If Moses Lake level

(1044 feet A.T.) were raised to 1100 or 1150 feet A.T., these horns would lengthen northward, and more horns would appear, as shown in Figure 4. None of them, however, would reach across the Soap Lake rim. At 1150 feet, one would reach up Ephrata channel, and, if flooded to 1200 feet, that channel would also have tributary horns.

If only those hairpin contour loops indicating southward-sloping surfaces are drawn, from the Bureau's topographic map R4-5720 (as ∧s in Fig. 4), the entire gravel fill traversed by the three channels is seen to possess this characteristic. Elongated and subparallel closed basins in them are common. Whether the grooves are separated by constructional forms, or erosion under sheets of water across the entire plain has made them, or both processes have operated, these northward-tapering channels or grooves cannot be explained as post-gravel dissection by local runoff. They must have been made under glacial water flowing southeastward across the basin.

Figure 4 could have been drawn with about half as many similarly oriented ∧s selected to show the very reverse topography, that of prows pointing northwestward. The combination of prows and grooves appears to make a good case for erosional origin under a widespread glacial torrent, thereby requiring a pre-existing unaccented gravel plain. But structure cannot be ignored in interpreting the origin of these forms. An unparalleled opportunity to study structure was afforded by the 175 miles of canal excavation completed in 1952 and the many large gravel pits made in development of the Columbia Basin Project.

Boulders, mostly basalt, are embarassingly abundant in some pits, requiring excavators to discard a large portion of the gravel, and are very obvious in spoil heaps along canals. Columnar outlines are common, the greenish patina of a preglacial weathering rind surviving in large part on plane or concave sides, and fresh basalt showing chiefly on bruised and battered projecting edges. Most of them are distributed sporadically in cobble or pebble gravel, its constituents usually not much worn but without weathered surfaces. Many boulders may have been carried by floating ice, but the boulder-laden lee of scabland knobs and buttes in the basin indicate the operation of some agency capable of tearing boulders loose. One large gravel pit (Sec. 17, T. 20 N., R. 29 E.) shows two scabland knobs at the bottom. The large quantity of rejected boulders in the pit came almost entirely from the basal part of the 30-foot section. Everywhere that the bottom of the gravel is known along the two deeper channels, the older sediments are lacking, and the gravel rests directly on unweathered scabland basalt.

The common fore-set bedding of the basaltic gravel, containing erratic material and caliche fragments, is not ordinary current fore-setting in horizontal courses of limited thicknesses; it is more like deltaic fore-set structure extending through the entire exposure and deposited at the angle of stability. Only the higher, flatter tracts commonly have nearly horizontal bedding. The long fore-sets are found in excavations on steep slopes where they dip either in conformity with those slopes or at some angle between the slope direction and south. (Examples: the Willow Lake road cut approximately in sec. 6, T. 21 N., R. 28 E., the Larson Air Field pit in sec. 31, T. 20 N., R. 28 E., the Columbia Sand and Gravel Co's pit in sec. 5, T. 19 N., R. 28 E., a pit a mile north of Moses Lake town in sec. 11, T. 19 N., R. 28 E. and the huge pit in the south edge of sections 1 and 2, T. 17 N., R. 28 E.)

The Quincy basin gravel deposits are terraced only along the deeper portions of the Rocky Ford channel, and only the creek and lake indicate a meander pattern in any part of any channel. The notch of Rocky Ford channel in the Soap Lake gravel rim is 110 feet lower than that for Ephrata channel and 50 feet lower than the col at the head of Willow Springs channel. A post-flood glacial Grand Coulee river (a detoured Columbia) and/or discharge of a post-glacial lake in Lower Grand Coulee may well have been responsible for the only normal stream-valley pattern in all of Quincy basin's or, indeed, any other scabland gravel deposits.

The higher bottom portions of Rocky Ford, deepest and widest of the three channels, possess the same aberrant pattern of anastomosing minor channels, ∧-shaped heads of channel grooves, abundance of elongated closed basins

FIGURE 5.—NORTHERN END OF QUINCY BASIN
From U. S. Bureau of Reclamation topographic map R4-5720, sheet 1. Contour interval 10 feet.

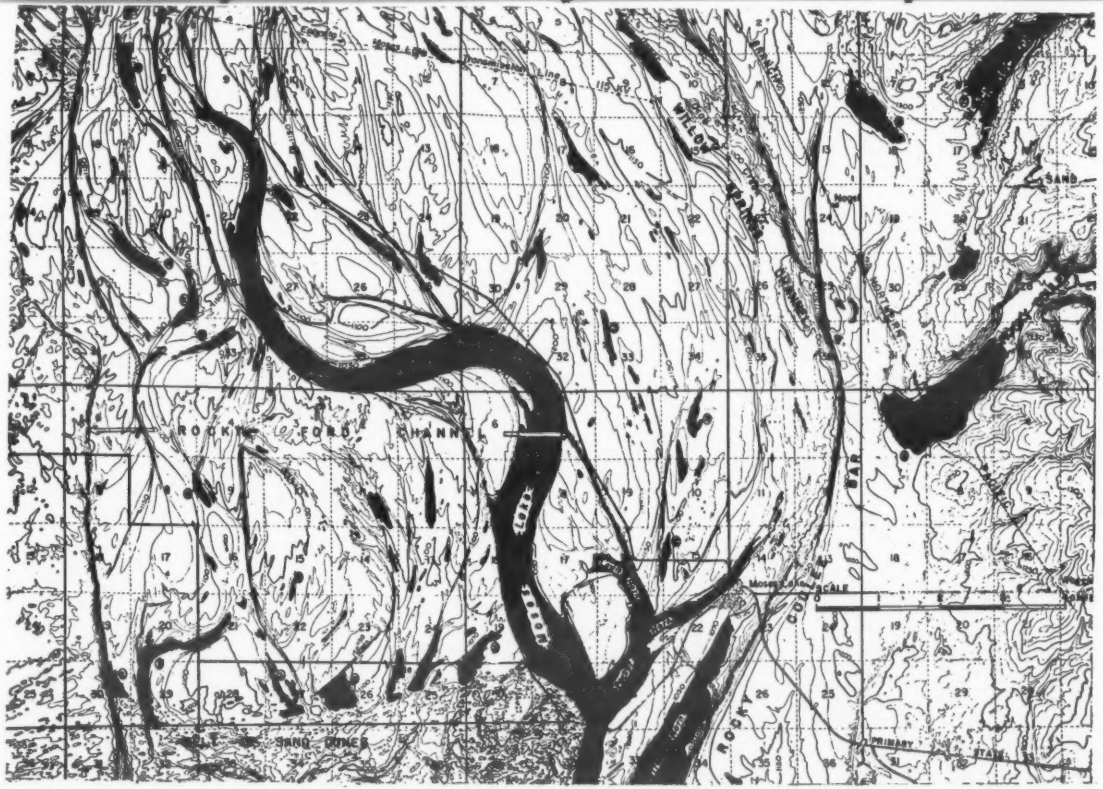


FIGURE 6.—MOSES LAKE AND VICINITY

Closed basins in black with number of hachured contours indicated by circled figures. Map shows: 1—constriction, partial closure, and deflection of Rocky Coulee by bar (with recurved fingers) at entrance to Quincy basin, 2—continuation of the bar southward past Moses Lake town, 3—drowned stream channel containing the lake, 4—dune belt which dams the channel, 5—A-shaped, flood-made anastomosing pattern of channel-like grooves up to 100 feet above the lake, 6—scarp on west side of Rocky Ford-Moses Lake channel rising to the higher gravel plain containing Ephrata channel.

Part of U. S. Bureau of Reclamation topographic map R4-5720, sheet 1. Contour interval 10 feet.

in those grooves, and boulder-strewn areas that Quincy basin has over most of its higher surfaces (Fig. 6).

Even if the three channels across Quincy basin are sequential, as is suggested by their differing altitudes, the channeling must have started with all three, unless we use an implausible, unrecorded rotating sequence of ice dams. They were made under a flood of adequate volume to spread initially over the entire gravel surface, a flood well recorded in the horns of Ephrata channel. But this western channel was too far offside to be deepened as much as the other two, hence was not occupied by later floods that used and deepened Rocky Ford and Willow Springs channels. Because Willow Springs channel enters Rocky Ford at accordant grade, constituting the depressions for the horns of Moses Lake, these two channels must have functioned contemporaneously to the end of the glacial flooding, even though the col of the head of Willow Springs is 50 feet higher than that of Rocky Ford. The need for large volume is again obvious unless we can believe that unrecorded ice jams in proper sequence determined the succession.

Two major points remain to be cleared up: the genetic conditions for the gravel fill into which the three channels were trenched, and the reason for this trenching across a tract of earlier glacial-river deposition.

West of Ephrata channel, a 10-mile width of unchanneled plain, underlain by basaltic sand and granule gravel only, lies 25–75 feet higher than the Ephrata channel head. It may be a remnant of preflood outwash once filling almost the entire width of Quincy basin. But east of Ephrata channel, no remnants of such a plain survive from the first episode of trenching. The two upland tracts separating the three channels differ in altitude by 60 feet. Both have the Λ-shaped groove heads on their southern slopes, undrained depressions on their summits, and long gentle slopes except where later trenching has encroached on them. The western upland has a subsidiary summit 70 feet lower and more strikingly streamlined. The writers can explain these broad hills only as having been subfluvially molded and in no way owing their forms to subaerial erosion. But uncertainty remains as to what extent they were actually built up where they stand.

The reason for the trenching in Quincy basin's glacial river deposits by later rivers of the same character is found in the southeastern outlet to the basin, Drumheller Channels, and is treated in the following section.

PLATE 3.—UNNAMED COULEE AND ASSOCIATED GRAVEL DEPOSITS

FIGURE 1.—Vertical aerial photo showing gravel deposits. Top is north. Shadows fall to the left. Southeastern slope of Trail Lake anticline can be traced diagonally across the photo, from the lower left, by means of the northwest-facing cliffs. The Λ-shaped small scabland hills in midlength of the coulee are also eroded portions of southeast-dipping basalt flows. The high gravel remnant is scarped on all sides, the eastern and highest scarp having been cut through the gravel and the underlying loess whose caliche cap is shown by the white line along the scarp. Two successively younger gravel deposits, with convex tops, lying southwest of the high gravel remnant, are bars, left by later floods, each becoming scarped as the outlets to Bacon synclinal valley were deepened. The cataract alcoves all head in the same flat-lying flow and presumably date from the last flood. Compare with Plate 2. Photographs AAR. 9A-9 and 11, Agr. Adj. Adm.

FIGURE 2.—Oblique aerial photograph, showing unchanneled scabland leading to the northernmost cataract alcoves. Shadows fall to the right. Looking northeast. Photo by H. T. U. Smith.

PLATE 4.—GIANT CURRENT-RIPPLE MARKS

FIGURE 1.—Two miles east of Odessa. Known to the writers only from the Production and Marketing Administration's aerial photograph AAO-66-247 (herewith reproduced) and from descriptions of the fire north of the road by Lincoln Co. officials. Scaled on the photograph, these giant current ripple marks have wave lengths of about 250 feet. A considerable part of the glacial stream here was a Cheney-Palouse distributary.

FIGURE 2.—Channel floor south of Warden cataracts. When these ripple marks were made, Drumheller carried a glacial river 9 miles wide. Photo by Ag. Adj. Adm. AAH 4A-50.

FIGURE 1

FIGURE 2

UNNAMED COULEE AND ASSOCIATED GRAVEL DEPOSITS

FIGURE 1

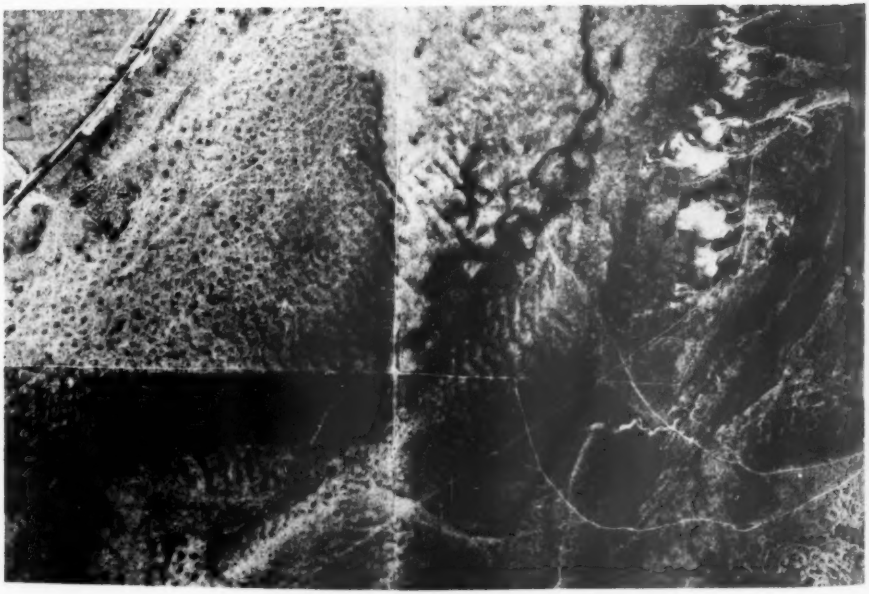


FIGURE 2

GIANT CURRENT-RIPPLE MARKS

FIGURE 1

FIGURE 2

BARS IN UPPER CRAB CREEK VALLEY

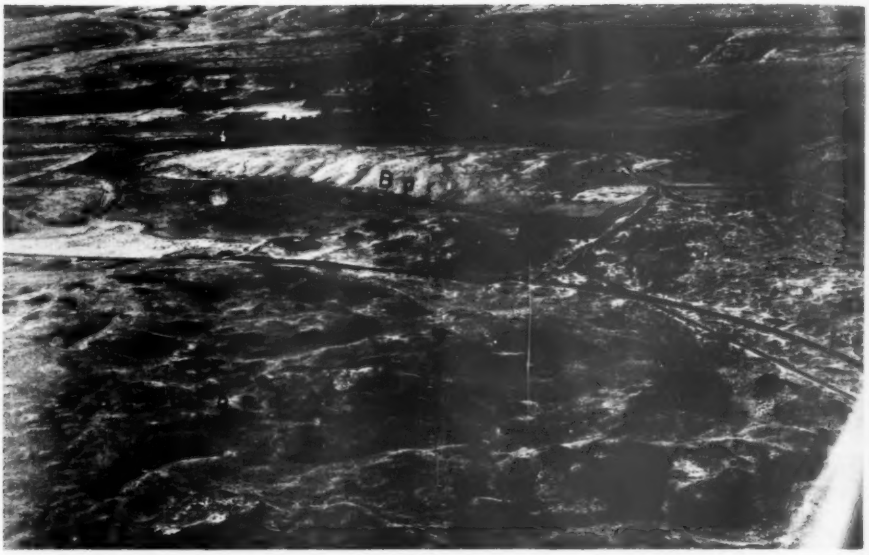


FIGURE 1

FIGURE 2

BARS IN UPPER CRAB CREEK VALLEY

Drumheller Channels

Maps: Corfu and Othello quads., U.S.G.S.; R4-5720, sheet 2, U.S.B.R.

(Bretz, Smith, and Neff)

This tract of scabland (Pl. 7; Pl. 10, fig. 1), the deepest dischargeway out of Quincy basin, is an almost unbelievable labyrinth of rock-walled canyons, basins, and buttes about 9 miles wide and 12 miles long, eroded as much as 300 feet deep in the folded basalt of the Lind flexure and the Frenchman Hills anticline. Its magnitude and complexity cannot be explained without the concept of enormous discharge across it from Quincy basin, but its history is more complicated than has ever been outlined. It appears to record at least three times of channeling in bedrock unprecedented except in Washington scabland.

Folded with the basalt flows of the Frenchman Hills anticline was an overlying, weak, sedimentary formation. Portions of it have survived all later erosion and crop out low on the anticlinal flanks. Subsequent to deformation and much subaerial erosion came deposition of later Ringold sediments, still flat lying.[10] It now appears certain that the site of Drumheller channels carried such a cover when the first glacial-river discharge out of the closed Quincy basin occurred and that the earlier channels were superimposed from that cover.[11] Both east and west margins of the channeled tract are scarps in sedimentary rock, in places more than 100 feet high. One surviving hill of this rock, 50 feet high and similarly scarped, stands about 2 miles out in the complex. In the 12 miles of Drumheller's length, bases of these scarps descend from 1300 to 950 feet A.T. on the east side and to about 800 feet A.T. on the west side. Although water may have lapped up somewhat on these scarps, the much greater weakness of the sedimentary rock than that of the underlying basalt indicates that scarps were undoubtedly eroded back and channels correspondingly widened until the contact of the two formations was essentially at the edge of the torrent. By this assumption, the surface gradient of water

[10] Neff considers Low Gap (alt. 1510 feet A.T.) in Frenchman Hills anticline (sec. 13, T. 17 N., R. 14 E., Beverly quadrangle) to be a wind gap recording drainage out of an early Quincy basin. This view requires two post-Ellensburg uplifts separated by a long interval of erosion during which the early fold was reduced to a mature topography. The interpretation is based largely on the existence of 138 feet of sand, clay, and caliche in the floor of the gap. Its basalt floor is about 175 feet above the surface of the Quincy basin fill a mile directly north, and approximately 375 feet below the Hills summit a mile to the west. The lack of stream gravel in the gap is taken to record an interval of subsidence and sedimentation following the erosion but preceding the second uplift which made the closed basin the glacial floods found. By this view, the sediments in the gap are early Ringold. Culver (1937) thought that some "Ringold" was older than the folding of Frenchman Hills.

[11] Suggested by J. P. Buwalda but, in error, rejected by Bretz (1928c, p. 218).

Plate 5.—BARS IN UPPER CRAB CREEK VALLEY

Figure 1.—Wilson Creek valley floor a mile above junction with Crab Creek. Looking southwest. Grain field crossed by road may be identified in Figure 1 of Plate 9. Photo by H. T. U. Smith.

Figure 2.—Bar No. 1, looking southward. Vegetation pattern outlines the current-ripple marks. Fosse, with cultivated field, lies between bar top and foreground scabland. Another cultivated tract, in lower left, is in the blocked mouth of a prescabland gulch marked VB in Plate 6. Bar without a number on Plate 6 lies in angle between Wilson Creek and Crab Creek, and its flood-eroded, 150-foot scarp faces observer. Not visible in photograph are giant current-ripple marks on summit of this bar. Photo by H. T. U. Smith.

Plate 6.—BARS IN UPPER CRAB CREEK VALLEY

Bar No. 2 depends from scabland at upstream end, tapers out in midvalley. Undrained fosse on south side is nearly as wide as creek-occupied valley bottom beyond the bar. Unrecognized until seen from the air are the giant current-ripple marks on this bar. Photos by H. T. U. Smith.

Figure 1.—Looking northwest

Figure 2.—Looking west

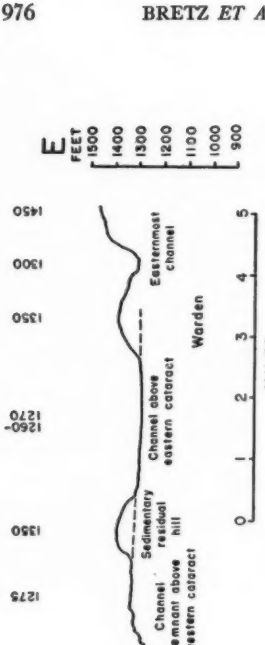


FIGURE 7.—PROFILES ACROSS DRUMHELLER CHANNELS.

Looking north. Profile follows a slightly zigzag course across the head of Drumheller Channels, between Warden and the nose of Frenchman Hills anticline. The easternmost channel has been but slightly deepened or scarped, and the Warden cataract channel floors have been cut but little into basalt. The three deep notches, by one of which Crab Creek escapes from Quincy basin, were made simultaneously. No preglacial drainage course across the anticline could have been deeper in basalt than the 1100-foot flat between Crab Creek and the two rock basins.

escaping down the southern slope of the anticline reached the surprising figure of 30–50 feet per mile. Yet the volume of the first discharge, which removed the overlying sediments, was sufficient to demand a width of 9 miles between bounding scarps. Drumheller at this earliest stage was an extraordinarily wide cascade over the edges of the tilted basalt flows.

Some basalt also was eroded during this first episode. There are two abandoned cataracts and recessional gorges in the eastern part of the group, 3–4 miles southwest of Warden (Othello quadrangle), one on either side of the isolated hill of sedimentary rock. One cataract brink is 1270 feet A.T., the other, 1250. Each is about 50 feet high. The western one has lost its western portion because of subsequent deepening to make rock basins 200 feet lower in the basalt. At comparable altitude and with similar height is part of such a cataract at the scarp base on the west side of the complex, its eastern portion destroyed by later channeling. With a very minor channel just east of Warden, they are the only records of the first glacial stream erosion in basalt at Drumheller Channels. (Fig. 7; Pl. 7.)

A significant item in understanding the mechanics of scabland cataract retreat in the vertically jointed basalt flows is the existence of a dike with irregular but nearly horizontal columns holding up the eastern Warden cataract lip. Plucking was not favored here, retreat stopped at the dike, and no horseshoe outline of the lip developed. Drumheller's great deepening occurred west of this dike.

The 300 feet of deepening since these Warden cataracts operated is assigned to the same later episodes that saw Bacon syncline notched after its high gravel was deposited, and Quincy basin's early gravel deposit channeled. This later deepening at Drumheller caused the reversal of behavior of glacial water traversing Quincy basin. The Warden cataracts are relics of an early scabland floor in Drumheller, probably correlative with the highest gravel in the Soap Lake rim. Sequence in channels is well shown in sections 10 and 13, T. 16 N., R. 29 E., where the channel below the Warden cataract is cut off by, and hangs 100 feet above, a later and deeper channeling. The floor of the hang-

ing channel shows giant current-ripple marks (Pl. 4, fig. 2).

A dam (O'Sullivan) 3½ miles long, constructed by the Bureau of Reclamation across the entrance to the lowest of the Drumheller Channels, will impound water up to 1052 feet A.T. and thus raise the level of Moses Lake 8 feet. A huge gravel pit in the south part of sections 1 and 2, T. 17 N., R. 28 E. was studied before it was submerged under this reservoir. The pit is in the flat above the eastern scarp of the united Rocky Ford and Willow Springs channels, a scarp subsequently cut into by local drainage from Lind Coulee.

The gravel is mostly of well-worn pebbles but contains cobbles and boulders up to 2½ feet in diameter, some sporadically distributed, others aggregated in strata. Caliche pebbles and granules are well distributed but not conspicuous constituents. The structure throughout the 40-foot section is that of a delta, with long uninterrupted, south-dipping fore-sets capped by a few feet of top-sets. Altitude of the flat is approximately 1085 feet, 100 feet above the near-by lowest channel floors leading to Drumheller. The scarp is as clearly erosional here as it is in the latitude of Moses Lake. The delta deposit therefore is the southernmost portion of that upper flat and antedates latest use of both the Rocky Ford and Willow Springs channels in gravel and erosion of the deepest of the Drumheller Channels in basalt. The 1070 contact of top-sets on fore-sets records the approximate bottom at the time of deposition. It was about 200 feet lower than the initial discharge over Warden cataracts. This 200 feet of erosion could have occurred during the first discharge southward out of Quincy basin, but more probably the delta is a later overlay (say, at Ephrata channel time) on the southern portion of Quincy basin's earliest and highest gravel, with the contact as yet unrecognized.

Whatever its age, there was a basin here to receive the delta. The basalt of Lind flexure dips below all trenching at this place, but the sedimentary cover must once have been present. Only flood erosion of a channel-bottom hole could have removed this weak rock before Drumheller was notched below about 1060 feet. Known thickness of gravel in this basin is 128 feet. Thus the deeper, later channeling in Drumheller, contemporaneous with Rocky Ford and Willow Springs late channeling, amounted to more than 100 feet.

Attempts to establish a profile northward from the Warden cataracts to the highest gravel in northern Quincy basin have not been satisfactory because terraced surfaces are lacking. A correlation, however, seems logical, and, if correct, it appears that a large proportion of the scabland gravel fill was made during the first flood down Grand Coulee, Crab Creek, and other glacial tributaries to the basin.

If this be correct, then either the spectacular erosion in Grand Coulee was consummated early, or great quantities of basaltic debris went across and out of Quincy basin during later floods. That the second alternative is correct is strongly suggested by the amazing number of enormous granitic boulders along the Rocky Ford and Willow Springs channels, some of them 10, 15, even 20 feet in diameter. The granite, far less conspicuous in the higher, older gravel and in the earlier abandoned Ephrata channel, presumably came down Grand Coulee after its great upper cataract had retreated past Steamboat Rock and uncovered the sub-basalt granite hill tops now so conspicuous in the head of the great coulee (Bretz, 1932, Figs. 33–37).

Upper Crab Creek

Maps: R4-5720, sheet 1 and G-5883, U.S.B.R.

(Bretz, Smith, and Neff)

General Statement

Several glacial rivers entered Quincy basin from the east (Pl. 1). The largest was Upper Crab Creek, leading southwestward across the northern divide from the preglacial Spokane and Columbia valleys. Smaller but highly significant ones were Rocky, Bowers-Weber, and Lind coulees, distributaries from the large Cheney-Palouse tract.

Crab Creek carried the discharge of seven glacial rivers from the northeast and five spill-over distributaries from the Cheney-Palouse tract. Although it already had a valley deep in basalt, at some time Crab Creek had

two broad, uncanyoned, spillover distributaries across the upland south of it and leading to the east side of Quincy basin. One theory (Flint, 1938) calls for so much aggradation by glacial streams in preglacial valleys that water spilled across interfluve tracts. Later removal of most of such fill has largely restored capacity of the original valleys. Upper Crab Creek valley contains great gravel deposits which, by that theory, are but terraced remnants of such a fill. An excellent place to test this idea is the vicinity of Wilson Creek's junction with Crab Creek, located where the supposed filling produced the distributary overflow.

Plate 8 is a 2-foot interval contour map of 12 sections in T. 22 N., R. 29 E., made by the Bureau of Reclamation. The scabland geology on it was mapped by Bretz and Smith. Bretz (1928a; 1928b) has published descriptions and figures of these deposits, identifying them as river bars so greatly exceeding the magnitude of known river bars that some alternative seemed called for. But this extraordinarily revealing map and the accompanying aerial photographs should make clear the impossibility of defending any alternative.

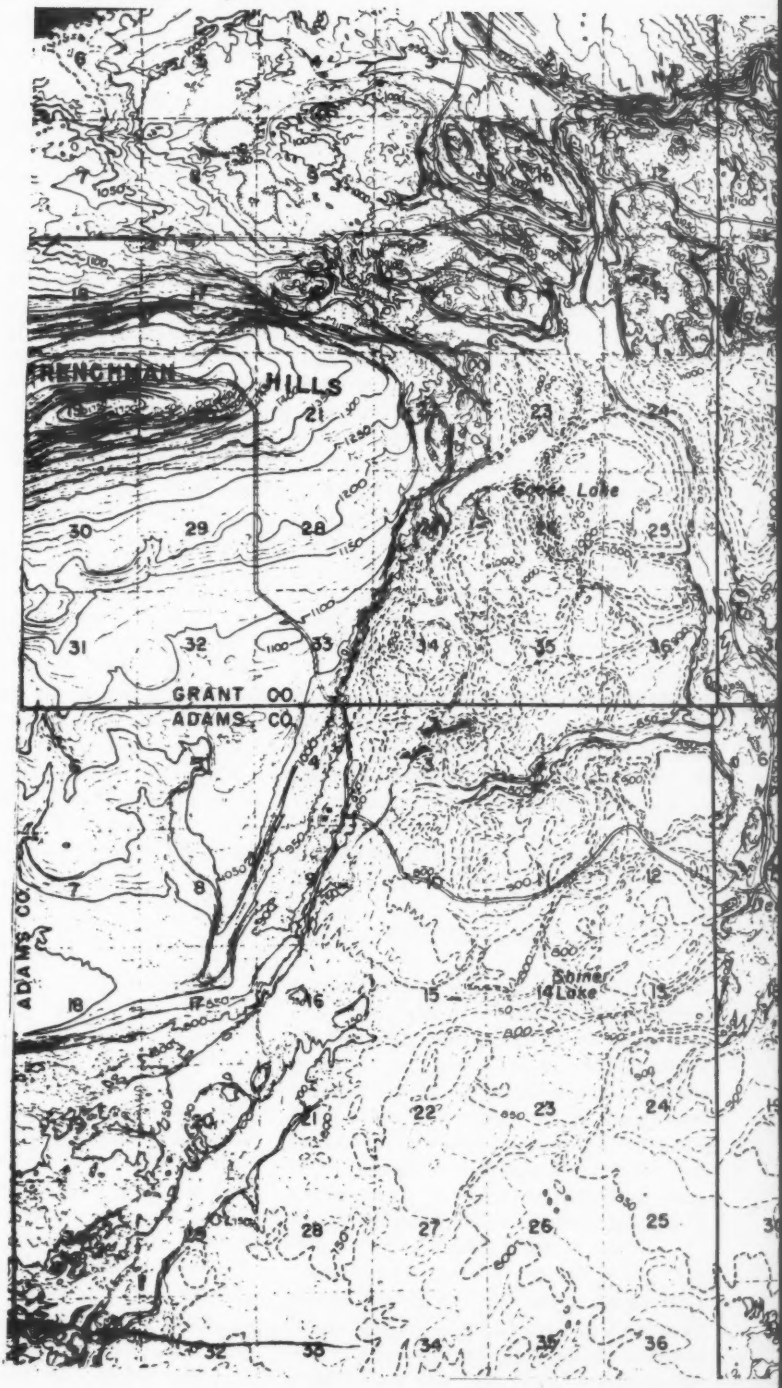
Bar No. 1, directly north of the town of Wilson Creek, has a broad summit 120 feet above valley bottom with linear undulations transverse to valley length and averaging 10 feet high and 250 feet apart (Pl. 5, fig. 2). The upstream (eastward) slopes of the low ridges are between 7° and 9°, the downstream slopes are 19°–22°. No excavations exist. Only medium-sized pebbles lie on the surface. Aerial photographs show the pattern of these asymmetrical undulations to be that of current ripples, weakest at the ends farthest from the center of the valley, strongest on the brink of the descent toward that valley. Two shallow undrained sags lie between the ripple-marked top of the bar and the scabland wall north of it. This bar tapers and lowers toward the west, continuing to hug the north wall of Crab Creek valley but having another shallow undrained sag on its north side. Two capacious ravines in basalt were blocked by the bar's growth, and subsequent drainage has cut much narrower ravines across the gravel blockade.

Bar No. 2 (Pls. 6, 8) lies along the south side of the creek valley opposite the western end of Bar No. 1. It can be well seen from State Highway 7. It extends westward from the scabland ledge carrying the highway, lying almost wholly out on the valley floor. The fosse between it and the ledge covers about 40 acres and was a closed depression 10–14 feet deep before the railroad grading.

A gravel pit low on the north slope of the bar shows uninterrupted deltalike fore-sets dipping diagonally down the slope and down valley. Another pit, at the extreme east end, close to the base of the basalt cliff from which the deposit depends, shows an anticline-like structure of steep, long fore-sets for the full height of this end of the bar. Highly angular basalt boulders up to 8 feet in diameter lie almost where they fell from the downstream-facing cliff into the growing gravel deposit. This bar summit is more than 60 feet above the fosse on the south and nearly 80 feet above the flat valley bottom on the north. The downstream tapering end is a quarter of a mile out in the valley from the nearest scabland wall.

Bar No. 3 (Pl. 8), also on the south side of the valley, shares with No. 2 in enclosing the fosse back of and between them, and rises 94 feet above that depression in a quarter of a mile. It is made irregular by a scabland shoulder which projects northward into it and rises above it as a knob of bare rock. Just beyond that is a small fosse in the bar summit. Thence the bar tapers down to valley bottom with another fosse, an eighth of a mile wide, between it and the scabland of the valley's south side.

A large shallow pit on the bar summit shows no structure but reveals many large, angular basalt boulders, 5–6 feet in diameter, and one 6-foot granite boulder. The remainder of the deposit is composed of pebble and small-cobble gravel, fairly well worn, with almost no sand or granule gravel. The nearest upstream source for basalt boulders is a knobby scabland bench 3000–4000 feet distant, its summit no higher than that of the bar. If this gravel hill is a constructional form, these boulders had to cross the fosse and then be rolled *up* the stoss end of the bar to the summit, nearly 100 feet higher. If it be an erosional remnant, and a stream bed such as Flint has postulated once existed for them to travel on from the scabland knobs, boulders 5 and 6 feet in diameter



MAP OF

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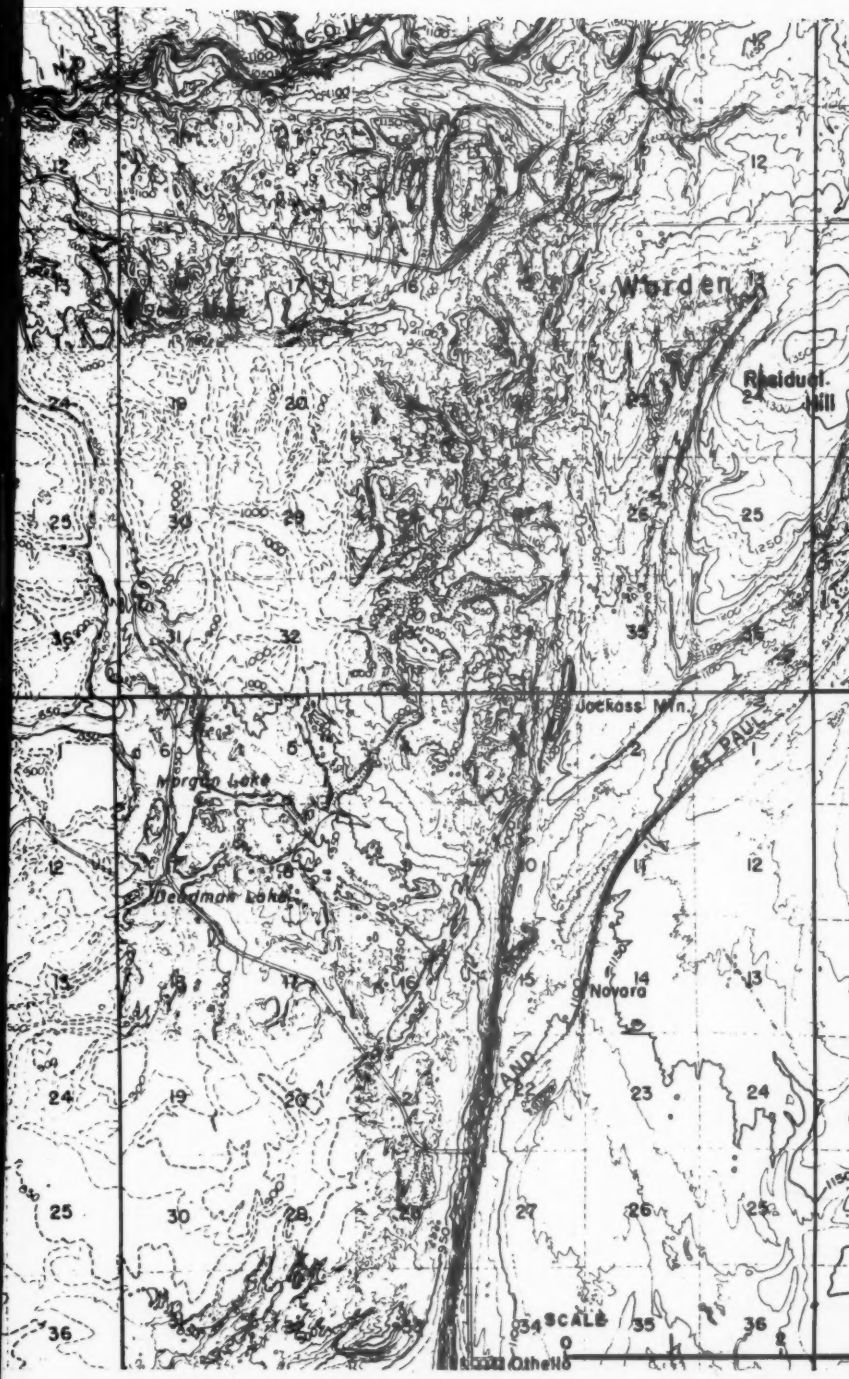
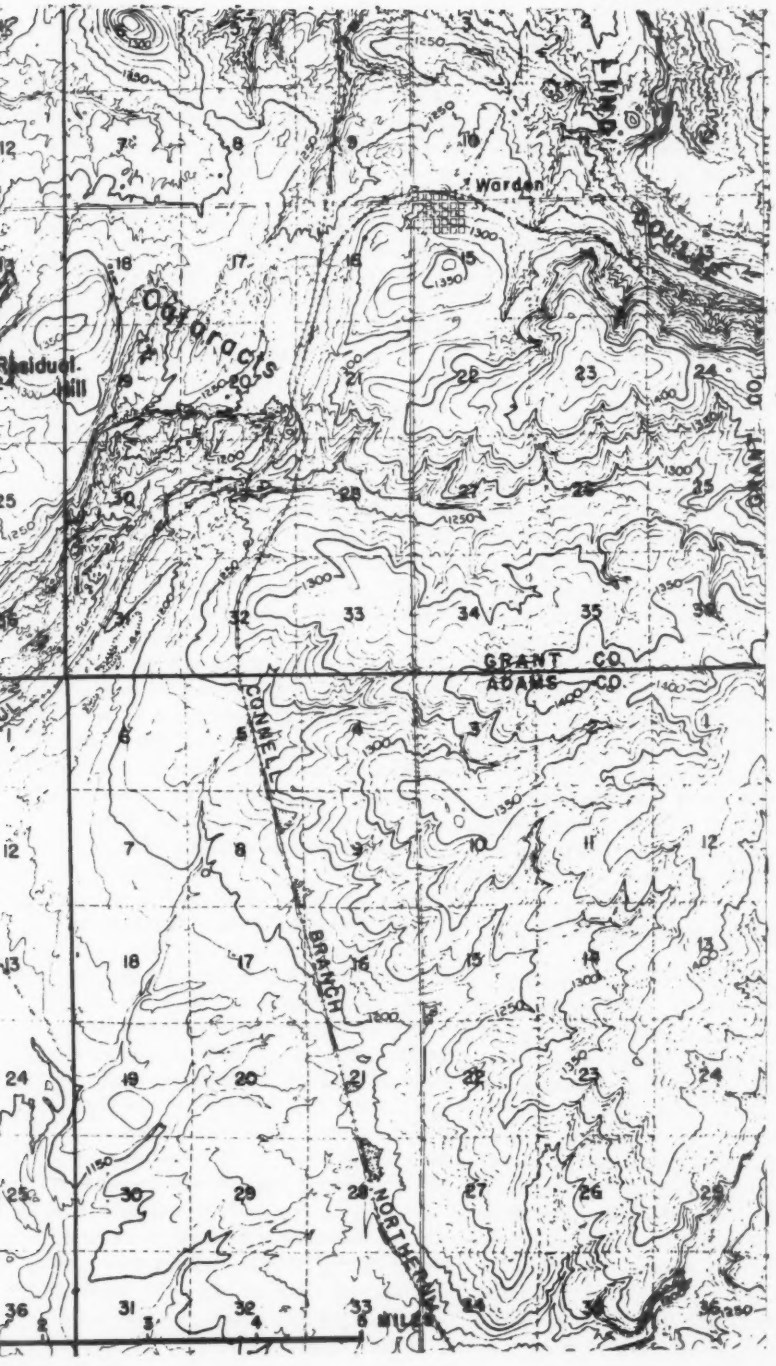


MAP OF DRUMHELLER CHANNELS, WASHINGTON

s contours 10 feet; for dashed contours 25 feet. U. S. Bureau of Reclamation topogr



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tion topographic map R4—5720, sheet 1.

moved more than half a mile on no gradient at all. Transportation by floating ice must be invoked unless the glacial river had a velocity competent to move them either on a level or even upgrade.

The valley bottom between Bars No. 3 and No. 4 (Pl. 8) is reduced to only a third or a fourth of its average width in this township. This narrows, confined only by gravel, is inexplicable by the fill-and-dissection theory. Even the empty fosse south of the west end of Bar No. 3 is nearly twice as wide as the narrows occupied by the creek. The outline of the prescabland valley is readily recognized from Plate 8, and the narrows immediately becomes understandable if Bar No. 3's distal taper grew downstream from an anchorage on the scabland shoulder in its mid-length. Cross-sectional room for the central bottom current of the glacial river was provided on the proximal end of Bar No. 4 where, over about 40 acres, the deposit grew only 20 feet above the valley bottom. A third of a mile farther down valley, the summit of Bar No. 4 is nearly 100 feet above valley bottom on either side. Length of the gravel hill and accompanying fosse is 2 miles. The streamless fosse to the north is nearly twice as wide as the creek-occupied part of the valley floor. Bar No. 4 lies in mid-valley, does not make contact with either side except at the very head.

This bar is defended by a splendid example of "trenched spur buttes" (Bretz, 1928b, p. 472, Fig. 23) where the tip of a preglacial divide between the main valley and a tributary from the north could not be tolerated by the glacial river and was plucked and notched by bottom currents into a linear group of ragged scabland buttes. The Great Northern Railroad traverses the length of the fosse and crosses this divide tip by one of the notches (Pl. 8). Only simultaneous functioning of all the notches could have produced or maintained the width of the fosse at its eastern end.

These streamlined bars assuredly are not terraces with outlines softened by subsequent slope wash. Their shapes and their fosse depressions cannot be due to leisurely erosion by late-glacial or postglacial streams. But may they not be remnants of a former valley fill which became gouged into the present topography beneath a flood or floods of the character read from Grand Coulee, Quincy basin, and Drumheller Channels? May not the bar forms be erosional instead of depositional? The composition and structure and the existence of fossae seem to debar that interpretation also.

These features do not fit Allison's ice-jam theory whereby properly located blockades in the glacial river laterally detoured the escaping water around the edges of the dams to make the scabland walls and to erode and rebuild remnants of an older, high-level gravel into bar forms. Allison's idea would better explain these features of Upper Crab Creek if his ice jam were placed somewhere upstream, and the alterations for which it was responsible were ascribed to a flooding downstream when the dam finally gave way. Judged by his interpretation of trenched spur buttes in Snake River canyon, he would make those of this tract either abandoned preglacial courses of Crab Creek or ice-jam "run-arounds."

But these Crab Creek bars are not high enough on the valley slopes for Allison's theory. They are not perched, their bases are essentially on the valley bottom, 100–200 feet lower than the bottoms of most of the notches. They do not contain any old gravel materials and are embarrassingly in the way wherever one may wish to place an ice jam. The features near Wilson Creek cannot be thus explained.

Comparison with Yakima Valley Gravel Deposits

A comparison of the topography, composition, and structure of these scabland gravel deposits with remnants of an extensive, deep gravel fill in Yakima Valley near Ellensburg are illuminating. Drainage areas of Yakima River above Ellensburg and of Upper Crab Creek are approximately the same. The Yakima Valley deposit is definitely older than Crab Creek gravels, but both, considered as once-continuous valley fills, have been subjected to erosion by glacial and postglacial water.

The Yakima gravel is terraced in two very definite levels. The lower terrace has as flat a surface as the modern flood plain. Erosion by local runoff is essentially nil, and the terrace is wasting away only from lateral attack by the river. This is the identical procedure of the

fill-and-cut theory. The upper terrace has a dissected margin of mature slopes but, back from the valleyward edge, is almost as flat as the younger terrace. There are no residual hills isolated on the upper terrace, and only one was seen on the lower. It stands so near the break between the two terraces and has such gentle slopes that it clearly is but an outlier of the upper flat.

Many highway cuts show the gravel of both terraces to be well sorted and well worn, far more so than most scabland gravel. There are some included sand lenses and strata of small boulders. Sorting and wear resemble that typical of scabland gravel only near basalt cliffs along the narrower part of Yakima Valley above Ellensburg. Here the gravel may be rather rubbly and contain large, angular boulders, obviously fallen from these cliffs during the aggradation. Stratification is horizontal almost everywhere, with only shallow fore-sets if any. Nothing resembling the long fore-sets of scabland gravel was seen.

The dendritic pattern of the slope separating the two terraces resembles nothing in scabland gravel topography. Dissection of the Yakima Valley fill here is a textbook example. Comparison of topography, composition, and structure with scabland gravel deposits is all contrast! Yakima Valley never had a Lake Missoula type of flood, and no barlike forms were made during deposition or dissection of its gravel.

Bars in Wilson Creek Valley

Bars up Wilson Creek Valley a short distance east of the mapped area carry even better-developed giant current ripples than those on Bar No. 1; maximum relief is nearly 20 feet. They occur at three levels, ranging from the broad valley bottom itself to the highest bar summit where they are close to the upper scabland, 150 feet above the valley floor. Ripples of the lowest group have wave lengths averaging 300 feet (Pl. 5, fig. 1; Pl. 9, fig. 1) and are most conspicuous where partially submerged in a flood-irrigated tract. Because two definite erosional scarps in gravel separate the three ripple-bearing surfaces, three floods are read from this topography. Except for cutting a narrow trench, postglacial Wilson Creek has done nothing to modify the valley floor, and existence of the ripples across the entire bottom proves that the last glacial river to use this route was not one of the leisurely meandering streams of Flint's concept.[12]

Giant current ripple marks are difficult to see at ground level, even aerial photographs fail to record them if the surface carries a good cover of sage brush (Pl. 4, fig. 1). The ripples shown cover a bar about 2 miles east of Odessa, some 25 miles up Crab Creek from Wilson Creek junction. There are 13 of them in a distance of 3300 feet. The clear depiction north of the straight east-west line, a road, is due to a brush fire shortly before the photograph was taken. Brush on the south side of the road was not burned, and the ripples show so faintly that they could well be overlooked.

The writers do not rest their case for enormous glacial rivers on current ripples alone but are convinced that only Pardee's interpretation (1942) for the Lake Missoula features will account for the ripple marks near Wilson Creek town.[13]

Black Rock Coulee

Reference has been made to two wide tracts of not markedly canyoned scabland in the downstream angle between Upper Crab Creek valley and Quincy basin, high above both lowland tracts (Fig. 8; Pl. 1). The southeastern of these divergences, Black Rock Coulee, takes off from the creek valley about midway between Marlin

[12] Flint came close to recognizing giant current ripples and thus to destroying any validity his hypothesis then appeared to possess. On page 477 (1938) he said "Here and there the fill surfaces [of his 'erosional remnants'] are marked by faint ridges up to several hundred feet long, 20 to 50 feet broad and a few feet high. It is quite possible that some of these are at least in part constructional. . . ." But "none of these forms, even if constructional, constistitutes, in itself evidence of deep streams".

[13] Subaqueous "sand waves" of this size have been known in the Mississippi River for more than 50 years (Lane and Eiden, 1940), in flood waters 80–90 feet deep. But both regressive and progressive sand waves fail to survive as flood velocity and depth decrease (Bucher, 1919). The features here described did survive, therefore are true current ripples. Unrippled original surfaces of scabland gravel deposits apparently did not experience a sufficiently high bottom velocity at close of deposition. An interesting sidelight on the hydraulics of these glacial rivers will appear when the giant current ripples are given careful detailed study.

FIGURE 8.—SUMMIT OF UPPER CRAB CREEK-QUINCY BASIN DIVIDE

Scabland, scarp, and surviving loess topography on summit of interfluve between Upper Crab Creek and east edge of Quincy Basin. Record of the early flood that entered Black Rock coulee. Altitude of highest and oldest current-rippled bar east of the town (*see* Pl. 9, fig. 1) falls within the altitude range of this scabland. Total width on summit is 5 miles, total vertical range about 200 feet. Heavy contours at 10-foot intervals. From U. S. Bureau of Reclamation topographic map R4-5720, sheet 1.

(Krupp) and Irby and enters Quincy basin about 12 miles south of the latitude of Soap Lake. Its altitude at the divergence is more than 250 feet above Crab Creek valley bottom, and at its junction with Willow Springs trench it hangs 200 feet. But Black Rock Coulee drainage does not enter the trench. About 3 miles short of the brink of the eastern Willow Springs scarp it turns southward and flows parallel to, but *on top of*, the scarp to enter the mouth of Rocky Coulee. In like manner, drainage from some 15 square miles of adjacent scabland north of Black Rock coulee is detoured southward without crossing the scarp (Fig. 5).

Black Rock Coulee debouches on a high basalt flat thinly covered with scabland gravel whose altitude strongly suggests contemporaneity with the high gravel separating the three channels in Quincy basin. Furthermore, this scabland between Crab Creek and the basin is not vigorously expressed, and some is weathered down into a coarse residual soil, unlike any basalt in the lower scabland of the region. Black Rock Coulee and its associated high scabland (Fig. 8) is probably an early flood record. The broad summits of the Crab Creek bars, streamlined in conformity with the valley shape, are higher than some of the rock basins and intricate channels of the scabland farthest west on the interfluve between the creek valley and Quincy basin, and are about as high as butte tops over most of that tract. Both the giant ripples and the scabland basins required adequately deep water for their production and must be contemporaneous. The bars are oriented for a deeper part of the glacial-river current, whose direction was determined by preglacial Crab Creek valley in basalt. In these relationships is convincing evidence that the interfluve overflow was not due to a Crab Creek valley fill or an ice jam at the time but resulted from a volume that demanded more capacity for its channel than preglacial drainage had ever made for its valley. Upper limit of scabland at the foot of loessial scarps on this interfluve is 1500 feet A.T. Floors of near-by valley-bottom fossae are at about 1250 feet. The glacial river was more than 200 feet deep at this place.

Crab Creek valley was dammed at its entrance into Quincy basin late in scabland history by a gravel deposit with eastward-(upvalley-) dipping fore-sets recording backfill from Dry Coulee (Fig. 5). Silts overlying the lower scabland gravel for 10 miles or more upstream show that the creek valley was ponded as a consequence of the bar building. The gravel dam, cut in section by State Highway 7, is at least 50–70 feet thick. Crab Creek has subsequently cut a narrow gulch through it. The backfilling episode must post-date all glacial-river discharge down Crab Creek. It is to be correlated with the last functioning of Dry Coulee, later than the deep notching in the west saddle of Bacon syncline. It must postdate completion of Upper Grand Coulee's recessional gorge and the final capture by Grand Coulee of all Cordilleran glacial drainage to cross the plateau.

Although constituting a flat-topped and still nearly complete valley fill, the eastward-dipping fore-sets of this Crab Creek deposit and the silts east of it debar its interpretation as a record of Flint's hypothetical aggradation. It is also impossible to explain as the consequence of any ice jam. The gravel was carried *upvalley* in a large glacial-river course.

With this valley-mouth dam is correlated a well-marked bar crossed by State Highway 7 a mile west of the scarp of the dam. It lies almost in the middle of Dry Coulee, its summit at about the same altitude as that of the dam. The bar top is 30 feet above the channel on the west, 50 feet above that on the east, and 80 feet above the bottom of a closed depression immediately upstream. A pit on its western slope shows long deltaic fore-sets dipping west toward the smaller of the two bounding channels.

Rocky, Bowers-Weber, and Lind Coulees

Maps: Moses Lake, Othello, Schrag, Wheeler quad. maps U.S.G.S.; R4-5720, 1 & 2, U.S.B.R.

(Bretz, Smith, and Neff)

The lower ends of these coulees are crossed by siphons of the East Low Canal (Pl. 1) where excavation and topographic mapping by the Bureau of Reclamation have revealed some instructive details of the scabland picture. All

three coulees are distributaries from the extensive Cheney-Palouse tract and in prescabland time all headed on the west side of a divide which was crossed during the scabland making. Flint (1938) made this divide crossing possible by aggrading the Cheney-Palouse tract with glacial outwash deposited under ordinary conditions, disputing Bretz who had earlier repeatedly called for enormous volume to explain the divide crossings.

The entrance of Rocky Coulee to Quincy basin is narrowed to an eighth of its width a mile farther upstream and about a fourth of its average width for the first 10 miles above the constriction. A closed depression occupies the coulee floor for the first 2 miles upstream. Its bottom is 100 feet *higher* than that of Willow Springs channel half a mile to the west, and its drainage (when it overflows), after passing the constriction, turns south and flows parallel to the Willow Springs channel, *but on top of the scarp*, nowhere more than 3 miles back from the brink, until the scarp dies out about 6 miles to the south. Even there, Rocky Coulee's drainage continues its lateral course, refusing to join the main drainage south out of Moses Lake for 5 more miles and finally entering Drumheller Channels as a part of Lind Coulee discharge. Thus, with Black Rock Coulee's detoured drainage there is a total of 18 miles of this anomalous southward flow closely parallel to the Willow Springs channel but well above it and nowhere descending its scarp.

The topographic map suggests, and field examination proves, that Rocky Coulee drainage, beyond the constricted entrance to Quincy basin, is kept from spilling over the Willow Springs channel scarp by a long, low barrier of scabland gravel which tops the scarp and slopes gently eastward away from it. This barrier is most pronounced 2 miles southeast of Moses Lake town where crossed by U. S. Highway 10. Figure 6 shows a form resembling a wide spit pointed south with fingers recurved eastward. The highway cut on its eastern (lee) edge shows only loose basaltic sand and granule gravel. No structure is revealed. On the flat immediately to the east, roadside and irrigation ditches show caliche *in situ* for several miles eastward, and only the lack of a loessial cover here suggests alteration by glacial water.

Directly east of Moses Lake town, a large pit in this barrier shows coarser material at 1190 feet A.T. with some boulders of basalt and caliche but no interbedded sand or silt. The conclusion that the barrier is a natural levee type of bar and that it grew laterally eastward on to a higher marginal surface seems inescapable. This feature which constricted and detoured Rocky and Black Rock coulees is a part of the high gravel plain, older than the deep channeling across the basin. Where crossed by Highway 10, it is 80–100 feet lower than the Warden cataract lips. Either it was built after Drumheller had been considerably deepened or it was deposited in water 75–100 feet deep. The bar form is readily identifiable for 6 or 7 miles.

The lower few miles of Bowers-Weber Coulee (Pl. 1) carries a silt deposit, well exposed in road cuts along U. S. Highway 10 east of the caliche-covered plain above-noted and conspicuous as far east as Schrag (1244 feet A.T., Pl. 1). The deposit consists of alternating silt units, one finer than the other, each perfectly sorted, each with crinkly laminae that look like weak current records. The finer-textured member is prevailingly thicker, reaches a maximum of 18 inches, and is lighter-colored. The coarser silt grades up into the finer. Berg-deposited lenses of coarse sand containing non-basalt pebbles and a granite boulder 15 inches long indicate deposition in glacial water.

These silts lie only a little lower than the Warden cataract lips and should be correlated with their functioning. Ponding north of and below the 1250 Warden sill seems to be recorded while nonflood glacial drainage was arriving in Quincy basin, perhaps during the closing stages of the first flood episode or the early stages of the second.

Lind Coulee is crossed by the East Low Canal 4 miles east of Warden. Two siphons are to be used, a scabland butte (the summit of Lind flexure) in mid-coulee carrying a short stretch of open canal between them. The south end of the southern siphon terminates in a gravel deposit (altitude 1247 feet A.T.) on the coulee slopes where excavations expose a 20-foot section soon to be obliterated by back filling.[14] The material is all fine gravel and coarse sand, per-

[14] Exploration for foundations here penetrated 88 feet of gravel.

haps 20 per cent caliche bits, the remainder bluish-black basalt, stratified in repeated courses of short fore-sets which without exception dip away from the open Lind Coulee and toward the adjacent, small, tributary valley. The topography here looks like a rounded-off terrace, and, at first glance, the structure and composition seem to support that interpretation.

But the key to understanding the deposit is in the consistently 5- to 10-degree coulee-ward dip of the different nearly horizontal courses. They dip parallel to the surface slope, each course inclined in the *opposite* direction to that of its component fore-sets. All strikes parallel that of the slope. The deposit is a typical river bar marginal to the central current, its accretions swept back up and out of the main current and added to the accumulation in a protected re-entrant. The bar originally dammed the minor tributary here and extended at least 3000 feet farther down-coulee, grading into finer material with less-marked stratification. It must date back to the functioning of Warden cataracts.

Twenty-five miles farther up the coulee and a mile east of the town of Lind, at the junction of two glacial rivers (Pl. 1), the floors of three pits in scabland gravel are so covered with rejected boulders that one may cross without stepping off boulders. They average 3 feet in diameter, a few are 6 feet. Almost all are clearly from large-columned basalt knobs in sight up the coulee. Among the bruised but unrounded, unsmoothed basalts was one angular granite, 3 feet in diameter.

At Ritzville (Pl. 1), near the divergence of one of these coulees from Cow Creek valley in the Cheney-Palouse tract, a poorly sorted, bouldery gravel lies on the floor half a mile downstream from a low, channel-bottom, basalt knob. The very little-worn boulders are fortuitously distributed in the deposit. Some are 5 feet or more in diameter. One 3-foot quartzite boulder was found. Poorly developed fore-sets dip down along the coulee.

At Ralston (Pl. 1), near the divergence of the other coulee from Cow Creek valley, is a similar showing of bouldery gravel on the channel floor, with fragments up to 3 feet in diameter. There are no basalt outcrops near by. Neither here nor at Lind or Ritzville is there any body of finer, well-sorted material. It is all coarse, comparable, except for the boulders, to that shown in Flint's Figure 2 of Plate 2.

Both Ralston and Ritzville tell the same story that was read at Lind. Glacial torrents capable of tearing large blocks of basalt loose and transporting them along the low gradients of the preglacial coulees left these records. Such bouldery stream debris derived and transported on such gradients (about 25 feet per mile) cannot be harmonized with Flint's theory of moderate streams.

Floors of the coulee heads near Ritzville and Ralston hang 50–75 feet above the bottom of Cow Creek valley, which is but a modified preglacial drainage way. By the hypothesis here advocated, the larger scabland tract was under water deep enough to reach the top of the former divides in loess. These boulders were dislodged from parent ledges by the distributary streams. In the absence of steep gradient, only great volume will suffice for the velocity indicated. A basalt bench at least 50 feet above the coulee floor at Lind carries scabland gravel. This high-lying, divide-crossing distributary should have been abandoned at the very beginning of Flint's dissection episode, and its 50-foot fill should still be intact.

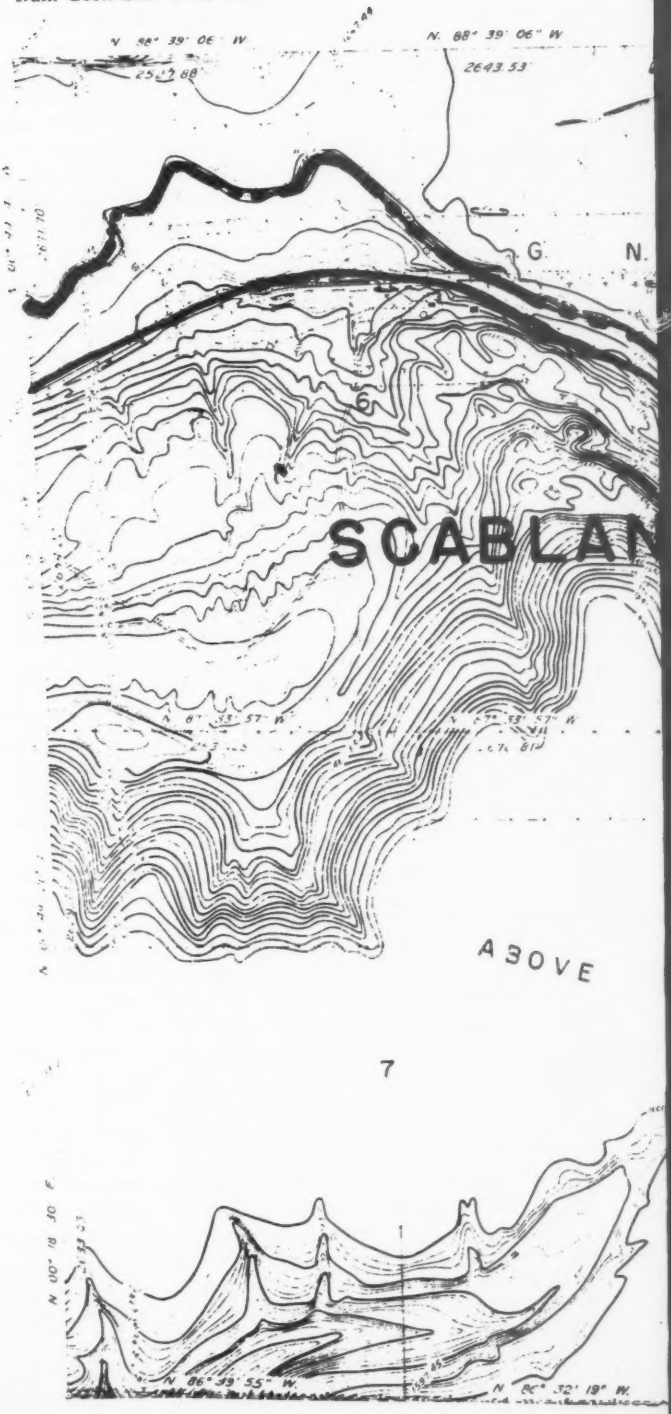
If several floods traversed the Cheney-Palouse tract, some deepening of Cow Creek's preglacial valley probably occurred. The evidence at the mouth of Bowers-Weber Coulee is that no flood came down it from the Cheney-Palouse tract after the silting episode, which is correlated with the first flood in Quincy basin. Therefore, these high-lying, distributary coulees between Crab Creek on the north and Washtucna Coulee on the south are believed to have been occupied only during an early scabland making. If so, removal of any hypothetical gravel fills must be dated from that same early episode, before the Bowers-Weber silts were deposited.

Western Part of Quincy Basin and Adjacent Columbia River Valley

Colockum Pass, Malaga, Quincy quad. maps, U.S.G.S. and U.S.B.R. map R4-5720, sheet 1

(Bretz, Smith, and Neff)

Babcock and Evergreen ridges constitute one long, low, north-south, structural elevation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SCABLAN
ABOVE
7
N. 86° 39' 55" W.
N. 86° 32' 19" W.

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N. 88° 39' 06" W.
VALLEY
G N.
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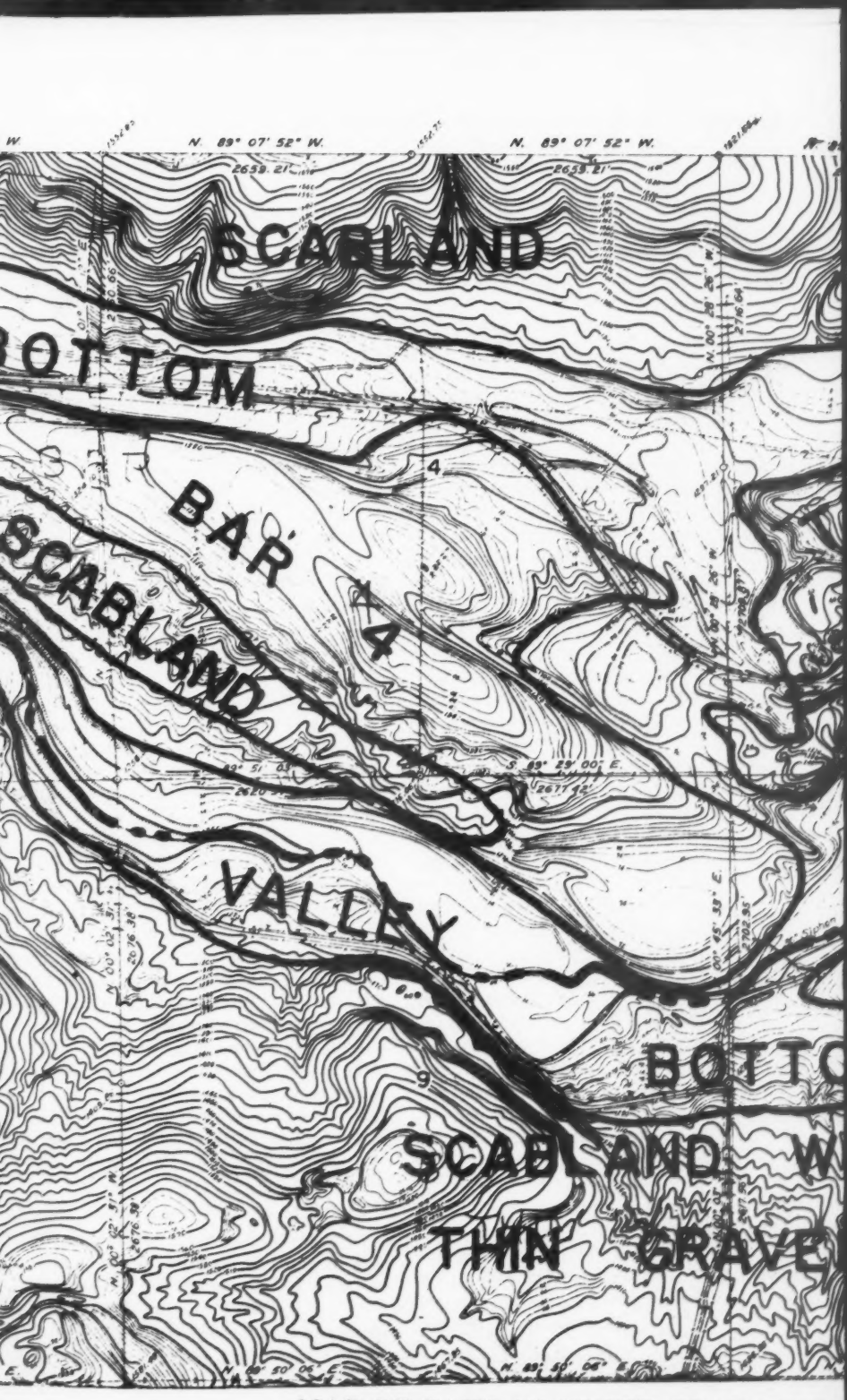
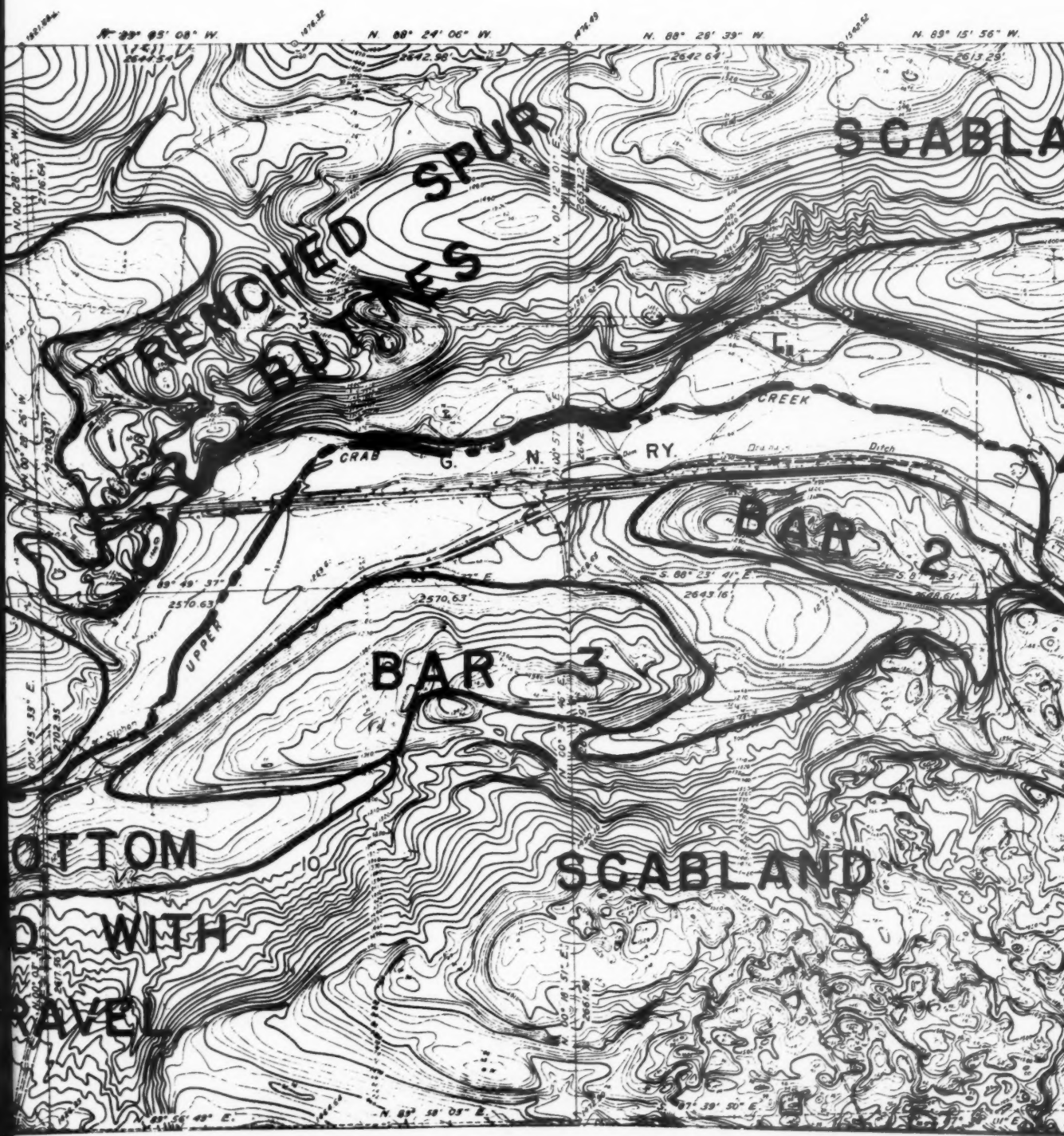
BLAND
1600'
8



MAP OF BARS IN UPPER CRAB CREE

Below junction of Wilson Creek glacial river. Part of U. S. Bureau of Recla



RAB CREEK VALLEY, WASHINGTON

. Bureau of Reclamation topographic map G 5883. Contour interval 2 feet.

BLAND
V.B.
BAR I
WILSON CREEK
VALLEY
BOTTOM
BAR
UPPER CRAB CREEK
12
N 88° 39' 25" W.
N 89° 21' 12" W.
N 89° 54' 48" E.
265.64
N. 89° 54' 58" E.

separating Quincy basin from Columbia River valley, cut into two by recession across it of the huge double Potholes cataract, nearly 1½ miles in total width. (*See* Pl. 1) The two ridges are actually the toe of the much greater uplift west of the Columbia (Table Mountain), amputated by the river in maintaining its antecedent course. Of several sags in the elongated summit of this ridge, two more (Crater and Frenchman Springs) also became dischargeways to the Columbia. Just west of each sag, the water plunged several hundred feet over the valley's eastern wall. Two of the cataracts thus engendered (Potholes and Frenchman Springs) receded essentially the full length of the notch across the ridge, but the third and smallest (Crater) retreated so little that its channel in the floor of its notch is still largely intact.

At the head of all three spillways, the Quincy basin surface slopes eastward away from Babcock and Evergreen, so that channel-bottom altitudes are readily obtained. Crater's channel head is 1280, the other two are 1210. But the original notches were so modified that estimates of initial altitudes found by the discharging water may vary with the observer. Bretz has long maintained that the cataracts were contemporaneous, with the upper limit of stream trenching in each notch at about 1300 (the same as at Warden cataracts), and that the 70-foot higher altitude of the channel to the smallest falls is the result of less volume and briefer use.

One of three conceivable alternatives to this idea of contemporaneous functioning is Meinzer's assumption of a remarkable coincidence in subsequent warping which brought the cataract upper limits into approximately the same horizontal plane. Another is an equally remarkable rotating succession of unrecorded ice jams. A third alternative is that, at the time of cataract discharge, the basin possessed an evenly graded alluvial fill leading from the mouth of Grand Coulee down to the three outlet places, and that shiftings of the detoured glacial (but nonflood) Columbia on this aggrading plain made possible a succession in use. Such an occurrence would be recorded in considerable remnants of the plain between adjacent spillways lying at the altitude of the highest channel scars and sloping up back into the basin.

But the slope of the plain, 10–20 feet to the mile, is down toward the southeast all the way to Ephrata channel and Frenchman Hills. It certainly is not the original slope of an outwash plain leading to the cataracts. Furthermore, it is largely unchanneled and in 25 50-foot U.S. B.R. test holes boulders have been found only close to the cataracts' channel heads. Its eastern part is heavily covered with basaltic sand and granule gravel, and the western part is composed mostly of sands and clays with a heavy caliche. Two caliches separated by sand or cemented gravel in some holes indicate prescabland origin. The cataract channel heads are cut into a caliche cap.

The eastern portion may be part of a normal outwash plain, but its slopes indicate drainage out through Drumheller before Ephrata channel time, not by way of the cataracts. Thus the falls may date from an early flood, after which ordinary meltwater discharge from Cordilleran ice may have built a plain in Quincy basin adjusted to Drumheller's depth at that time. Later floods then destroyed all but the extreme western part of this plain. Only the freshness of the cataract features (Pl. 10, fig. 2) seems out of harmony with this sequence.

A gravel deposit with most surprising structure lies on this plain a few miles southeast of the head of the Potholes, well exposed in a large pit (SW ¼ Sec. 32, T. 19 N., B. 24 E.) a mile east of the former post office of Burke. The material is coarse, with many basalt cobbles and some boulders whose diameters reach 5½ feet. Caliche boulders 3 feet in diameter are present, also several large granites. The stratification is dominantly highly inclined with dips toward the east, southeast, and south, *away from* the cataract head. Either these are backset beds as suggested by J. H. Mackin[15], or they record a reverse flow through the Potholes notch, out of the Columbia Valley and into the Quincy basin; Neff holds this latter view. Imbrication is so poorly developed that no decision as to which way the water flowed could be made from the deposit. If flow was eastward, the debris could have been obtained over the 3-mile stretch between The Potholes and the pit, at an altitude of about 1240 feet. A continuous gradient exists thence southeast and east to Drumheller's deepest channels. If the flow was westward, all debris must have been imported in floating ice.

[15] Personal communication.

It is conceivable that a flood down Columbia River, but not across the scablands, produced an ice jam in the narrows where the river crosses the Saddle Mountains anticline, 20 miles farther south (Pl. 1). However, such a jam would detour the Columbia flood not alone back through The Potholes notch but also back through Crater, Frenchman Springs, and Lower Crab Creek. Furthermore, the Saddle Mountains gap is 1¼ miles wide and only a mile long, and at present depth of the narrows the dam there would have to be 500 feet high. Detoured discharge would have to bypass a complete ice blockade by way of Othello Channels at the east end of Saddle Mountains. The lowest and largest of these channels has a floor at about 900 feet A.T. But it is difficult to believe that such a dam ever could exist. This problematical Burke gravel needs further study.

The steep, gullied, but largely soil-covered and caliche-bearing slopes descending into the Columbia Valley at the northern end of Babcock Ridge (Quincy and Colockum Pass quadrangles) have one great cliffed alcove of far more recent origin, recording the 200-foot spill of glacial water over Crater cataract. Superposition of this cliff and alcove on the older topography was accompanied by remaking of the gulch into a scabland canyon; Willow Springs Draw, whose floor strangely enough is an aggraded flat 950 feet A.T., hangs about 450 feet above the Columbia, 2½ miles from the cataract.

The flat is continued out in the Columbia valley as a terrace for 1–2 miles on both sides of the mouth of the Draw. Composition and structure are well exposed in the mouth where the terrace is a puzzling combination of buried

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PLATE 9.—GIANT CURRENT-RIPPLE MARKS

FIGURE 1.—Giant current-ripple marks at three levels in Wilson Creek valley a mile above junction with Crab Creek. Note how cultivation pattern along contact of grain field with valley bottom bar shows magnitude of ripples. The dark area is a flood-irrigated alfalfa field. Photo AAR-3A-80 of Agr. Adj. Adm.

FIGURE 2.—Staircase Rapids bar. Valley east of the ripple-marked surface was partially blocked by bar growth, and the tortuous stream meanders are on a silt flat. Drainage pattern east of creek is that of Palouse Hills topography. Production and Marketing Adm. photo AAP-26-188.

FIGURE 3.—Giant current ripple marks in Snake River valley at mouth of Devils Canyon. Production and Marketing Adm. photo AAR-9A-11.

PLATE 10.—DRUMHELLER AND POTHOLES DISCHARGEWAYS

FIGURE 1.—Drumheller Channels topography. Looking southward from O'Sullivan Dam. U. S. Bureau of Reclamation photo. Px222-117-142.

FIGURE 2.—Northern alcove of The Potholes cataract in 1954. Water is waste and leakage from West Canal about 2 miles distant and 75 feet higher than cataract brink.

Magnitudes involved: Lake in plunge basin is 125 feet deep. Top of bar on north side of alcove is 50–75 feet above lake surface, 175–200 feet above bottom of plunge pool. Bar fosse largely filled with talus. Cliffs back of bar are, including talus, 300 feet high. Scarp of gravel terrace along the Columbia is 150 feet high, and its summit is approximately 250 feet above river. Nearer cliff, across which lake discharges, is 150 feet high above gravel terrace.

Sequences involved: Cataract originated approximately at the near end of the plunge basin. Its total retreat was nearly 2 miles. Tip of the ridge separating the two alcoves is about an eighth of a mile back from the prescabland line of cliffs. Basalt bench below this cliff (part of bench south from Crater cataract) is gashed by glacial-river chutes, recording cascades rather than cataracts and presumed to be of subfluvial origin. Gravel of the terrace overlaps a little on scabland of this bench.

Correlations proposed: Gravel terrace has right altitude above the Columbia to be correlative with West Bar and Beverly Bar. The inconspicuous but recognizable, very bouldery, flattish mound buried under this gravel is a dump from early cataract retreat, comparable to that at Vantage Bridge. Scabland beneath terrace gravel is older than the terrace. Cascade chutes are later than the initial Potholes cataract. They and the bar north of the plunge basin may date from a late flood out of Quincy basin. U. S. Bureau of Reclamation photo P-222-177-35237

FIGURE 1

FIGURE 2

FIGURE 3

GIANT CURRENT-RIPPLE MARKS

FIGURE 1

FIGURE 2

DRUMHELLER AND POTHOLES DISCHARGEWAYS

FIGURE 1

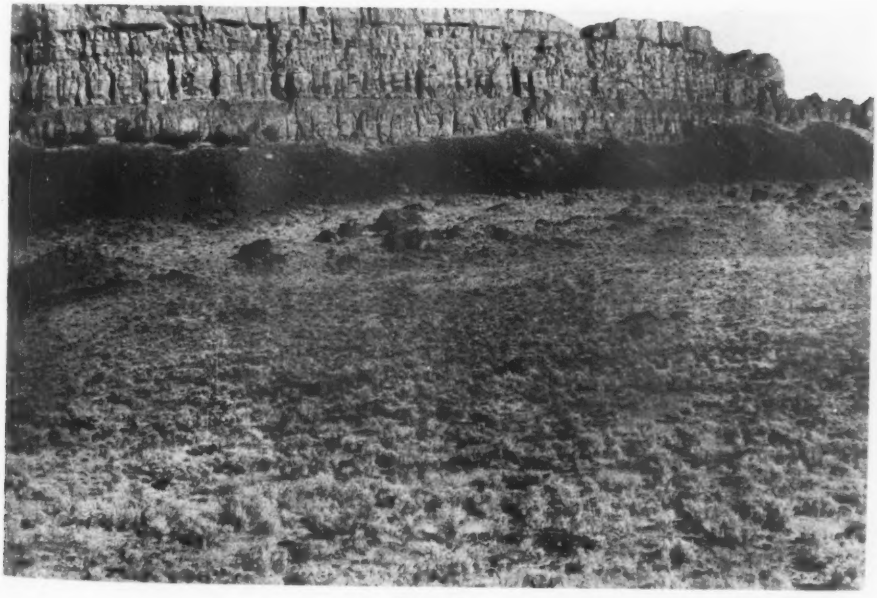

FIGURE 2

BENCH BETWEEN BABCOCK RIDGE AND COLUMBIA RIVER

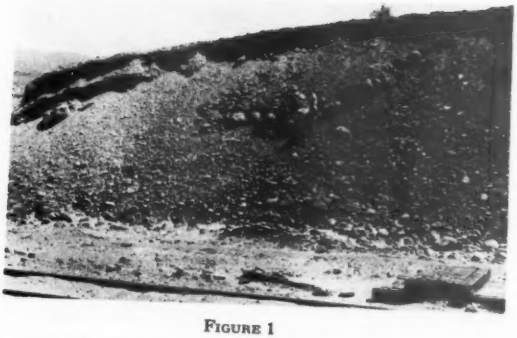

FIGURE 1

FIGURE 2

FIGURE 3

SCABLAND BARS

and semiburied scabland knobs, granitic sand, and scabland basaltic debris, into which postglacial drainage from the Draw has cut a narrow gulch and become superimposed on a buried shoulder of basalt.

The granitic sand, beneath the basaltic debris, is part of a Columbia valley fill not definitely identified farther down stream but conspicuous for miles upstream. There is but little basalt in its well-sorted, uniform material, and it has very few and only thin gravel members. It is a record of a leisurely Columbia's slow aggradation from river level, 550 feet A.T., up to rude terrace tops at 1300 feet A.T. Remnants back in tributary gulches testify to the existence of a valley system of present proportions when it was laid down, clearly in prescabland time. Apparently its dissection was also well advanced when the first flood discharged over Crater cataract.

In contrast is the coarse, very poorly sorted, almost 100 per cent basaltic debris in the same 950-foot terrace, strongly fore-set out of Willow Springs Draw, cobbly and bouldery and unworn to an unusual degree even for scabland torrent deposits. The terrace top does not appear to record a Columbia valley bottom at that time; it seems better interpreted as a later, lateral beveling. The fact that Crater has no empty plunge-pool basin, in contrast with Potholes and Frenchman Springs, strongly suggests that deepening of the two larger notches so decreased Crater's volume that its earlier plunge basin could not be kept flushed out.

The cliffed western side of Babcock and Evergreen ridges is the eastern wall of the Columbia River valley. It carries a great rock bench, locally a mile wide, extending from The Crater southward past both The Potholes and Frenchman Springs cataracts for the full length of the Quincy quadrangle (Pl. 13, fig. 2). The surface of this bench descends from about 1350 feet A.T. at the north to 875 at the south—a structural slope. Its topography is very instructive. Above 1300–1350 feet, the bench has gently rolling surfaces without outcrops, and the cliff overlooking it is only a smoothly sloped, steep hillside, also lacking in outcrops. There is no scabland in the view.

South of this, as far approximately as the crossing of the 1200-foot contour (Quincy quadrangle map), the bench is definitely scabland with buttes and rock basins but very few actual rock outcrops; almost all surfaces are thinly covered with weathered rubble. The cliff above it is talus-covered almost or quite to the summit, and the talus has a soil and vegetation cover (Pl. 11, fig. 1).

South of the 1200-foot contour, the bench is typical, vigorously expressed butte-and-basin scabland, and the cliff is bare, actively growing talus climbing only half to three fourths of its height (Pl. 11, fig. 2).

The bench is held up by the same basalt flow throughout the range of these contrasts, and the only material obscuring any part of its surface is a gravel bar overlapping the southern part of the more subdued scabland and possessing a fosse 20–30 feet deep between it and the cliff talus. It is a lateral accompaniment of the fresher topography. Here is scabland of two different ages.

A small deposit of gravel belonging to the older scabland lies in the west end of a notch across Babcock Ridge (sec. 20, T. 20 N., R. 23 E.) which was not reached by glacial water

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PLATE 11.—BENCH BETWEEN BABCOCK RIDGE AND COLUMBIA RIVER

FIGURE 1.—Cliff above older scabland. Looking northeast. Photo by H. T. U. Smith

FIGURE 2.—Cliff above younger scabland. Looking southeast. Photo by H. T. U. Smith

PLATE 12.—SCABLAND BARS

FIGURE 1. C.M. and St. P.R.R. section through Beverly bar, showing foreset bedding composed of boulders. Looking south. Man in photo for scale.

FIGURE 2.—Near Sperry. Foresets dipping into coulee from higher marginal scabland.

FIGURE 3.—North of Harder. Terminal part of "terrace." Skyline and railroad mark horizontality in the view.

from Quincy basin. It is a bar with convex surface (Pl. 13, fig. 2) about 50 feet above the scabland bench and it blocks the mouth of the notch and detours drainage past it on both sides. A granite cobble was found near the top. Another such deposit has a gravel-pit exposure (near the line between sections 17 and 18) between 1350 and 1375 feet A.T. on the nonscabland northern part of the bench. This deposit is composed largely of mixed basaltic granule gravel and granitic sand but contains considerable rubbly cobble basalt; the fragments are completely angular and much weathered. One fresh granite boulder was found. The pit is on the slope of a small sag (a little below 1350 feet) which became a minor, high-level channel across the nonscabland tract at the height of the earlier flooding of the bench. The locally derived coarse material was already loose, weathered debris, and the granitic sand and bluish-black granule gravel were carried here by the current that bore the berg-floated granite.

The southward slope of the structural bench is about 30 feet to the mile, and its entire surface below 1350 feet was unquestionably swept by the first flood. Did the cataracts out of Quincy basin function at that time?

If the surface gradient of the flooded Columbia was 10 feet to the mile, the cliff now notched by Potholes cataract must have been nearly submerged, and Frenchman Springs cataract could have had no more than 50–60 feet of plunge. A lesser gradient would bring the flood surface even closer to the brink of this southern cataract. Under such conditions, Crater could not have existed. Indeed, if there was no comparable flooding of Quincy basin at that time, the Crater notch must have carried water east into the basin.[16] The cataracts apparently belong to the second episode of flooding when, after a long interval of weathering, older scabland below 1200 feet on the bench was refreshed.

That there was a debris-empty Columbia valley of present depth when the western cataracts operated is clear from exposures made in relocating U. S. Highway 10 from Vantage Bridge to the summit of Frenchman Springs cataract. One now can see most of this splendid, abandoned waterfall from the window of his car. The floor of the cataract alcove is about 875 feet A.T., the Columbia below it is a little above 500. Debris from the cataract making, dumped over into Columbia valley, extends down at least to 600 feet A.T. and is well exposed in the highway's long, diagonal grade across it. Perhaps the most outstanding collection of rejected boulders in any scabland gravel pit occurs at the head of this grade. Their average diameter is 3 feet, a few reach 6 feet. They have been rolled at least a mile from some large-columned flows in the cataract ledges, but many still retain columnar outlines with only bruised, battered, and chipped edges. There is almost no sorting in the pit or highway exposures and no aggregation of boulders into strata. What rude stratification does exist is that of long fore-sets dipping toward and down into Columbia Valley. A confirmed skeptic is likely to try to make this deposit a talus!

It may well be that no one short-lived flood could have made all these cataract scars. Certainly no flood of the magnitude recorded could be more than short-lived. However, it is almost certain that the cataracts were contemporaneous, and surely no river of moderate size is recorded by the double Potholes cataract, 1½ miles wide. The debris spilled into the Columbia at Frenchman Springs is not a record of many centuries of slow cataract retreat. The way the lip of Potholes breaks into a series of smaller chutes (AAR-8F-194 and 195, also U.S.B.R. 10-foot contour map) is evidence that the last discharge over this cataract was of lessened volume. This may record the shrinking last stages of a flood or a later, smaller flood. But it was the last discharge over Frenchman Springs' southern cataract that dumped the great rubble into a debris-empty Columbia Valley.

Moses Coulee

Maps: Malaga quad., U.S.G.S.

(Bretz and Neff)

Moses Coulee, second only to Grand Coulee in magnitude, enters Columbia valley only a

[16] Since the field study by the three authors, Neff has found a mounded bouldery deposit exposed in West Canal sections at the head of the Crater cataract channel (SE¼ Sec. 14, T. 20 N., R. 23 E.) which contains slightly bruised columnar fragments 7 feet long in a largely unsorted gravel of caliche, basalt, interflow siltstone, and intact masses of loess. He feels that this deposit records an eastward flow through the Crater notch at some time previous to the last discharge westward.

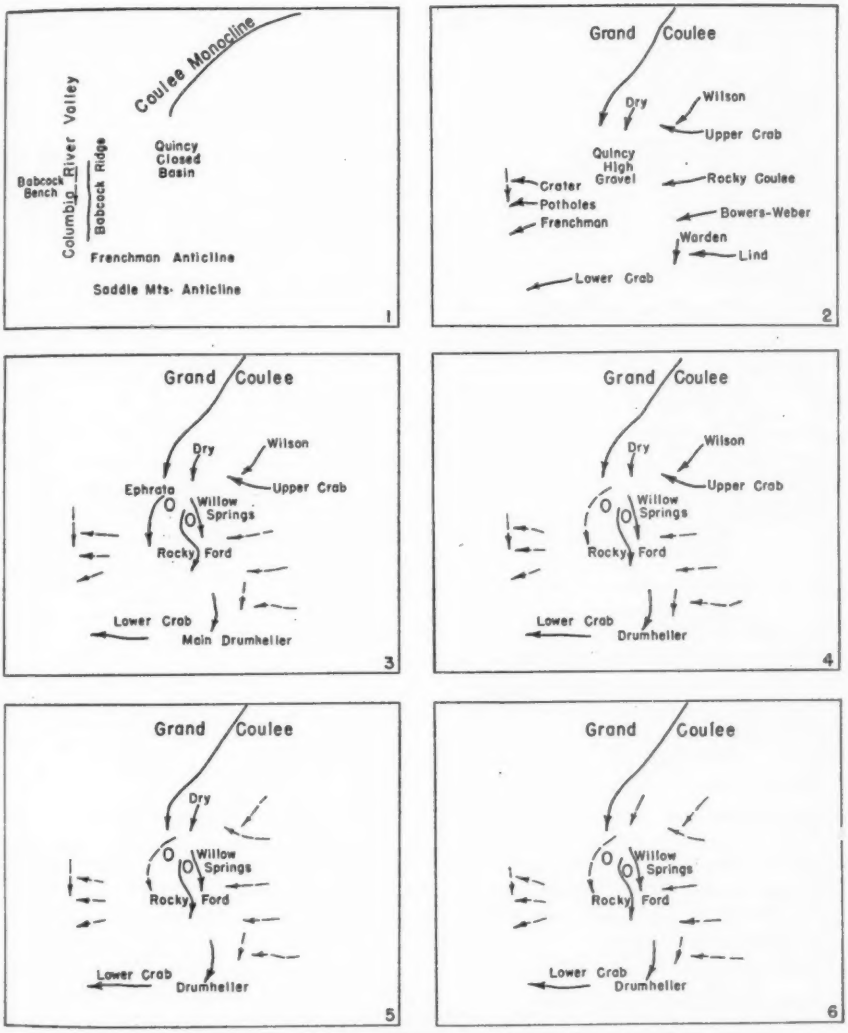


FIGURE 9.—SEQUENCE IN CHANNEL OPERATION, QUINCY BASIN AND ENVIRONS

few miles upstream from Crater cataract (Pl. 1). Although its headward connections with the rest of the plateau scabland complex are largely concealed by deposits of the latest Okanogan lobe to cross the Columbia, there is fair evidence of a scabland surface at about 2400 feet A.T. beneath the drift as far south as the coulee head, but no farther. The well-developed terminal moraine of this lobe blocks Moses Coulee at the crossing (10 miles from the head) with hills half as high as the coulee is deep, and associated outwash gravel extends down-coulee for a few miles in valley-train fashion. The great coulee, therefore, already existed in essentially full development before this latest glacial transgression on the plateau and must predate the later events in Grand Coulee history.

At the debouchure of Moses Coulee into the Columbia valley are remains of an enormous deposit of very little-worn boulders, sand, and gravel, nearly all basalt; its fore-set structure indicates a once-complete blockade of the

greater valley. Presumed moraine topography just upstream from this blockade and on the same side of the river appears to record contemporaneous glacial ice in Columbia valley. Bretz (1930b, p. 384) has suggested that this was made by a Wenatchee Valley glacier pushing out into Columbia valley. But Page (1939) has since shown that any such glacier must date further back than the late Wisconsin advance of the Okanogan lobe, and its drift should show pronounced weathering, a feature it does not possess. Its debris, furthermore, is essentially all basalt. Perhaps it is a record of ice blocks floated down from a breakup of the Okanogan lobe. The early Columbia valley flood recorded on the bench west of Babcock Ridge probably entered the master valley via Moses Coulee.

Correlation of Events and Sequences thus far Presented

(Bretz, Smith, and Neff)

Inspection of Plate 1 and Figure 4 shows that, during and after the episode which saw Ephrata channel eroded, water had ceased to reach the three western cataracts. A provisional chronology from data thus far submitted therefore would include (Fig. 9):

(1) The earliest flooding on the Babcock bench, apparently before Grand Coulee had been initiated
(2) The earliest recorded plateau flooding (because of an early Okanogan lobe dam) with
  (a) Operation of Grand Coulee and Crab Creek glacial rivers,
  (b) Cataract making at the three western spillways,
  (c) Deposition of the high gravel of Quincy basin,
  (d) Operation of Black Rock, Rocky, Bowers-Weber, and Lind coulees,
  (e) Beginnings at Warden cataracts of the Drumheller spill across the nose of Frenchman Hills anticline.
(3) The flood which initiated the three channels across Quincy basin and today is recorded only by the Ephrata channel (the three western cataracts and the four eastern coulees then abandoned; Drumheller deepened)
(4) The flood which used only Rocky Ford and Willow Springs channels, and continued the deepening of Drumheller Channels (Upper Crab Creek still functioning)
(5) The last flood down Grand and Dry coulees, making the Upper Crab Creek dam and adjacent Dry coulee bar (Steamboat Falls destroyed, Bacon syncline's western notch left hanging, all scabland channels, east of Grand Coulee abandoned)
(6) The final flood down Grand Coulee alone which completed the great Dry Falls cataract and Soap Lake basin and used both the Willow Springs and Rocky Ford channels.

The vertical succession of bars in Wilson Creek valley should date from the second, third, and fourth floods, because deepening of the channels across the Quincy basin fill permitted trenching back up Crab Creek and because at stage 5 Crab Creek valley was blockaded by Dry Coulee gravel. A seventh flood is recorded by the Beverly bar at the junction of Lower Crab Creek and Columbia River.

Lower Crab Creek

Maps: Beverly, Boylston, Corfu, Othello, and Smyrna; U.S.G.S.; R4-5720, sheet 2; U.S.B.R.

(Bretz, Smith, and Neff)

Discharge across the nose of Frenchman Hills (site of Drumheller Channels) entered an east-west synclinal valley between this anticline on the north and Saddle Mountains anticline on the south (Pl. 1). The syncline deepens and broadens eastward from the Columbia to the Channels; descent of the surface of the basalt along its axis is about 250 feet in 27 miles. The Bureau of Reclamation terms this eastern low broadening the Othello basin. The scarps in superbasalt sediments bounding Drumheller Channels continue westward along the syncline for some miles and make it clear that Othello basin was still largely filled when the earliest scabland flood arrived. The top of the basalt lay so low in the center of the basin that it escaped the terrible treatment meted out to that upraised in the Channels area. There is a plane tract of several square miles in the bottom

of the syncline near Corfu where the original basalt surface, although denuded, was not eroded.

Lower Crab Creek, after passing southward through Drumheller Channels' lowest gash, turns westward and follows this syncline, against the pitch, to the Columbia, descending about 100 feet in 25 miles. West of the Corfu flat, the syncline has well-developed scabland all the way. Some features of this glacial river channel deserve examination with reference to Allison's theory of ice jams.

One of these features is Natural Corral (Smyrna, formerly Red Rock, quadrangle). Another (shared by the Smyrna and Beverly quadrangles) may be called the Jericho coulee (Fig. 10). Both are accessory lateral canyons semiparallel to Crab Creek with looped courses back in higher scabland slopes along the north side of the synclinal valley. On the maps they look like detours for bypassing short, wide ice jams. But neither one has an eastern (upstream) connection with the main channel comparable to the remainder of the "run-around" canyon, and at neither place is there any narrowness, shallowness, or crookedness in the main valley favorable for initiating and maintaining an ice jam. Furthermore, the Natural Corral lateral canyon is only 3 miles long, allowing for only a short jam in Crab Creek, apparently too short, like Saddle Mountains gap, to provide the frictional locking of ice on ice and on shore to maintain a dam for the many years required if a moderately proportioned glacial river eroded the Corral canyon.

The Jericho lateral canyon (Fig. 10) is tandem, its total length of 9 miles almost completely broken in two in midlength by the disappearance there of the south retaining wall. Its floor at this breakdown is 60 feet above Crab Creek a mile to the south, yet its ephemeral drainage fails to find the gap and goes on west, parallel to Crab Creek, to end in a closed depression only 13 feet above the creek. This closure is due to Columbia River's Beverly bar, a late feature once entirely blocking the Crab Creek valley mouth, and dealt with in a later paragraph.

Two sequential ice jams therefore would be required for making the Jericho tandem canyon a "run-around". What should be the field evidence, other than mappable forms, to support the ice-jam theory? Allison (1941, p. 673) cited "The dozens of piles of iceberg-rafted till on the northeast flank of Rattlesnake Hills and in Cold Creek valley" as proof that great quantities of bergs reached the wide Pasco basin. They, with isolated berg "nests" and lone erratic boulders in so many places in the scablands, are irrefutable evidence for abundant floating glacial ice (*see* Bretz, 1930b, p. 410, Fig. 7) and, where closely spaced, are equally good evidence for the grounding of large numbers of these bergs. In Pasco basin, they obviously drifted ashore in a wide semiponded tract. If densely packed in a narrow valley, there should be at least as good a record.

With this criterion in mind, the two broadest and flattest butte tops enclosed by these looped lateral canyons were traversed. Not one erratic was found, nothing but bare or nearly bare basalt. Berg ice never made an ice jam along this scabland route. River ice, also used by Allison and favored by Lupher (1944, p. 1493) for damming and detouring glacial streams, would inevitably trap many of the abundant bergs, and a marginal record of this should survive.[17]

Both Natural Corral and Jericho lateral canyons carry good evidence that they are only enlargements of prescabland drainage ways already cut through the sedimentary rock into basalt. All the minor gulches entering them from Frenchmen Hills to the north have either hanging or strongly marked knickpoint junctions, but their gradients above the junctions, projected southward toward Crab Creek, find the basalt "island" butte tops too high. They were tributaries to west-flowing minor streamways north of and semiparallel to Crab Creek in the floor of the synclinal valley streamways ancestral to Corral and Jericho canyons.

These minor gulches were in existence when the highest scabland flood swept the superbasalt sedimentary off the top of the rock terrace they incise and covered their eastern walls with river-bar deposits while leaving their upstream-facing walls nearly bare basalt. Lower Crab Creek's lateral loops are prescabland local drainage lines which, modified by overwhelming flood discharge, had begun to grow into an

[17] To explain the Othello divergence by an ice jam in lower Crab requires a top to the dam 200–300 feet above these basalt "island" summits.

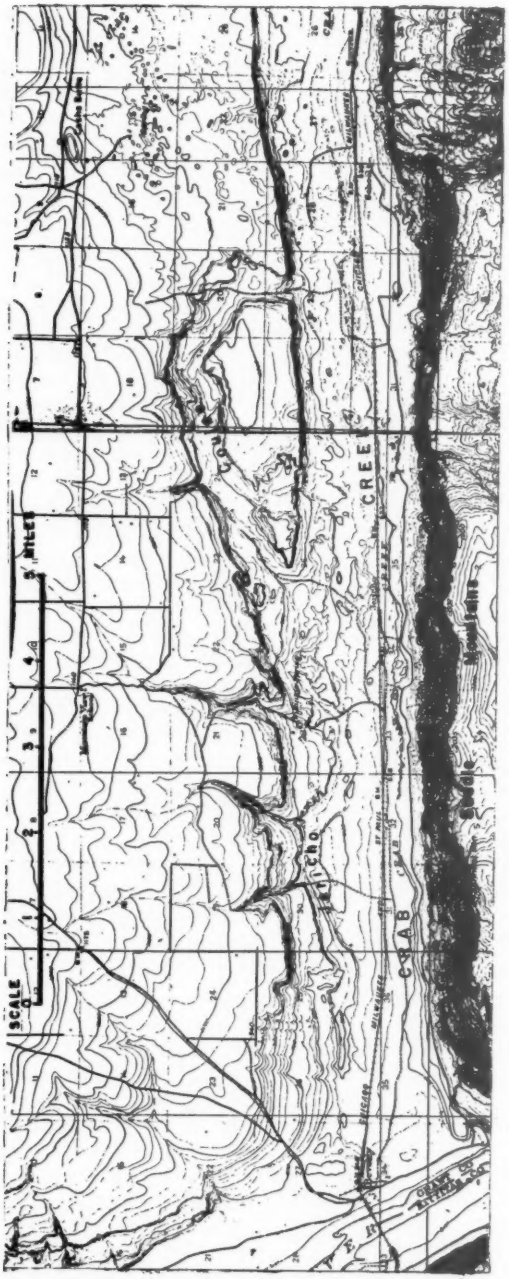


FIGURE 10.—LOWER CRAB CREEK AND JERICHO COULEE
Parts of Beverly and Smyrna topographic maps, U. S. Geol. Survey. Contour interval 25 feet.

anastomosis like that so well developed in Drumheller.

Perhaps the present course of Lower Crab Creek is, like that of Lower Grand Coulee, wholly of glacial-torrent excavation in the fractured basalt of the steep limb of a fold, and these lateral canyons out in the middle of the structural valley are modifications of the main preglacial drainage way.

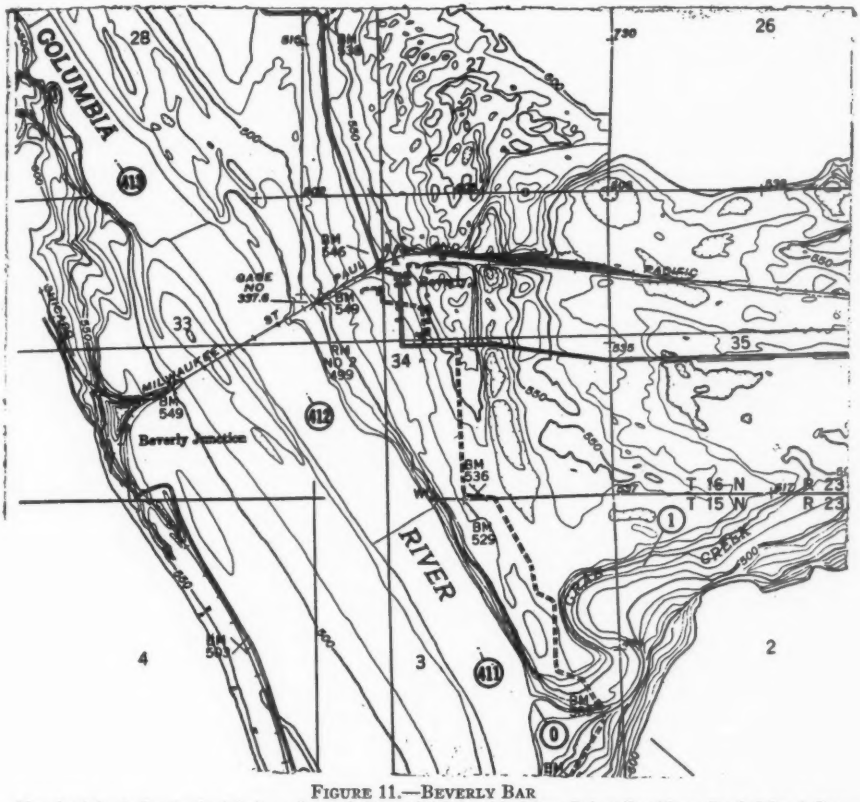


FIGURE 11.—BEVERLY BAR

Mouth of Crab Creek glacial river channel. Part of "Columbia River Priest Rapids to Rock Island Dam, Washington" map. U. S. Geol. Survey, U. S. Bureau of Reclamation and State of Washington. Contour interval on land 10 feet.

The Beverly bar (Fig. 11) is defended by scabland at the upstream angle between the Columbia River and Crab Creek and originally made a dam across the mouth of the creek valley. Its proximal end is 125 feet above the Columbia and almost 100 feet above the sump in which the Jericho lateral canyon terminates. Its linear form narrows and lowers southward for 1½ miles. It is actually made of two parallel bar ridges with a linear closed depression almost a mile long between them. Silts on the floor of Crab Creek valley prove that the creek was ponded by the bar. The creek has since cut a narrow ravine 40 feet deep across the low southern end of the deposit.

Composition and structure of the bar are well shown in a railroad cut nearly 30 feet deep across the higher, northern part. The western ridge is made of much coarser material than the eastern one. The riverward slope is almost wholly of boulders 3–6 feet in diameter. The body of this ridge shows long, lee (eastward) fore-sets, some of them almost entirely of boul-

ders up to 5 feet in diameter (Pl. 12, fig. 1). Other boulders lie embedded in cobble gravel. The debris in these deltalike beds, dipping east up Crab Creek valley, was rolled up the gentler Columbia slope of the bar and over the summit. Clearly the Columbia has been deep enough and swift enough at some time since Crab Creek ceased carrying a glacial river to throw this 100-foot barrier across the tributary's mouth.

The Beverly bar is nearly 30 miles downstream from the giant current ripples on West Bar (nearly opposite Crater cataract, Pl. 1; Bretz, 1930b, p. 400) whose altitudes range from 150 to 250 feet above the Columbia and up to 750 feet A.T. The giant ripples and the bar are probably contemporaneous.

The Beverly bar is 20 miles south of Potholes cataract, and its top is about 600 feet lower than the aberrant, east-dipping gravel near Burke. If the two gravel deposits, each recording a greatly flooded Columbia after the plateau glacial rivers had ceased to flow, were contemporaneous, and if the Columbia then had a surface gradient of 10 feet to the mile, water was 400 feet deep over the top of this bar. The two deposits are not necessarily of the same age, but there is no reasonable alternative explanation for the origin of the Beverly deposit. Saddle Mountains gap could not have been blocked when it was made. Because the bar was built after the last functioning of Lower Crab Creek glacial river, this Columbia flood must postdate the last of Grand Coulee's discharge. It is to be added to the floods previously listed, and perhaps to be correlated with the break-up of the Okanogan lobe dam described by Waters (1933, p. 763–820).

The morainelike topography in Columbia valley just upstream from remains of the Moses Coulee blockade is between 275 and 300 feet above the river, not too high to be correlated with the Beverly bar deposit and the West Bar current ripple marks.

One may best visualize the excessive erosion along Lower Crab Creek by comparing the character of Saddle Mountains and its adjacent syncline east of the Columbia (Beverly and Smyrna quadrangle maps) with that west of the Columbia (Boylston quadrangle map). The altitudes of the anticlinal summit and of the top of the basalt in the syncline on both sides of the river are about the same; the relief approximates 1300–1500 feet. Along Lower Crab Creek, glacial-stream erosion has bitten a channel into the basalt a mile wide on the bottom (far in excess of the creek's needs), whereas west of the river local drainage along the syncline has made only a narrow, V-shaped gulch (Johnson Creek) a third of a mile wide at the top.[18] The Chicago, Milwaukee and St. Paul railroad, crossing the Columbia 550 feet A.T. (bridge altitude), follows the floor of Lower Crab Creek channel eastward to climb only 200 feet in 20 miles. On the west side of the river, this railroad climbs nearly 500 feet (in a long diagonal) to get to the synclinal floor on basalt only 2 miles in a direct line from the river.

The steeply tilted and possibly faulted north face of the Saddle Mountains anticline west of the river is as well aligned as the east side. Rock outcrops and growing talus are almost completely lacking on this scarp. A cover of soil and vegetation mounts to the summit. Short gulches furrow the slope, but they show few if any outcrops. The debris-covered front descends with gradually decreasing slope, making a conspicuous concave profile. At the summit, this slope passes in most places without a break into a smooth, convex profile over the top.

East of the Columbia, the northern face of Saddle Mountains is cliffed above an actively growing talus which in places has mounted only half way to the summit, 1500–2000 feet above Crab Creek Channel floor. The talus extends to the bottom of this channel and has no concave profile. Locally, landslide is a marked feature of the front. A slide 4 miles wide between Corfu and Taunton (Corfu quadrangle) extends from the bottom of the scarp to the very top, 1¼ miles in the horizontal and 1200 feet vertically. There is no suggestion of slides west of the river; no undercutting by a great glacial river occurred there.

Although no field evidence is known to support the idea, the Taunton-Corfu slide may have been a factor in initiating the Othello Channels divergence (Pl. 1). If so, an under-

[18] Johnson Creek has been working through all post-deformation time. Lower Crab Creek, up to the time of diversion of glacial water down Grand Coulee, appears to have been scarcely larger than Johnson.

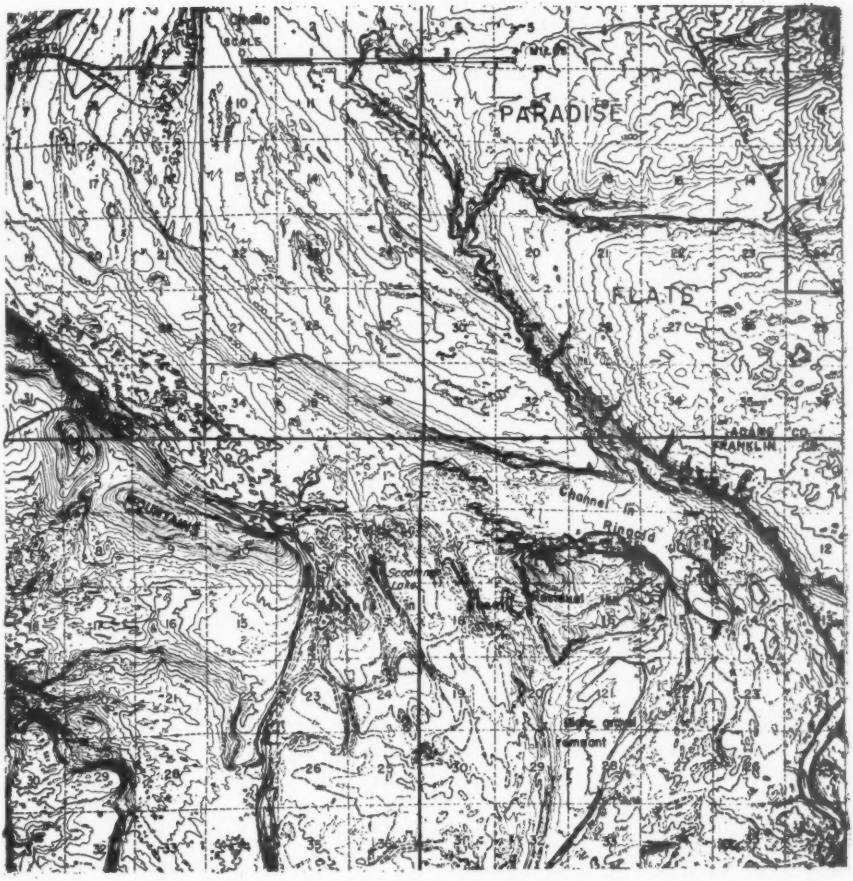


FIGURE 12.—MAP OF OTHELLO CHANNELS
Interval for continuous contours 10 feet, for dashed contours 25 feet. From U. S. Bureau of Reclamation topographic map R4-5720, sheet 2.

cutting of the mountain front by an earlier flood down Crab Creek alone seems required.

OTHELLO AND KOONTZ COULEE CHANNELS

Maps: Connell, Ellopia, Hanford, Othello and Scooteney Lake, U.S.G.S.; R4-5720, sheets 2 and 3; U.S.B.R.

(Bretz, Smith, and Neff)

Two groups of glacial-river channels lead south and southwest from Othello basin (Pl. 1); the Othello group (Fig. 12), crosses the axis of the Saddle Mountain anticline (much as Drumheller Channels cross the Frenchman Hills fold), and the distributary Koontz group (Fig. 17) crosses farther southwest, traversing a plain of sedimentary rock (Ringold formation) to reach the Columbia valley in the Pasco basin. They carried water from Quincy basin that failed to follow the Lower Crab Creek channel. The crossing was made at the eastern low terminal portion of the anticline and was initiated on a cover of sedimentary rock, in part folded but more largely of unfolded Ringold. Deep trenching into the deformed basalt flows beneath has made part of Othello Channels almost as wild

a scabland topography as Drumheller, although only a third as wide and a fourth as long.

The outstanding rock basin of Othello Channels, containing Scooteney Lake in the bottom

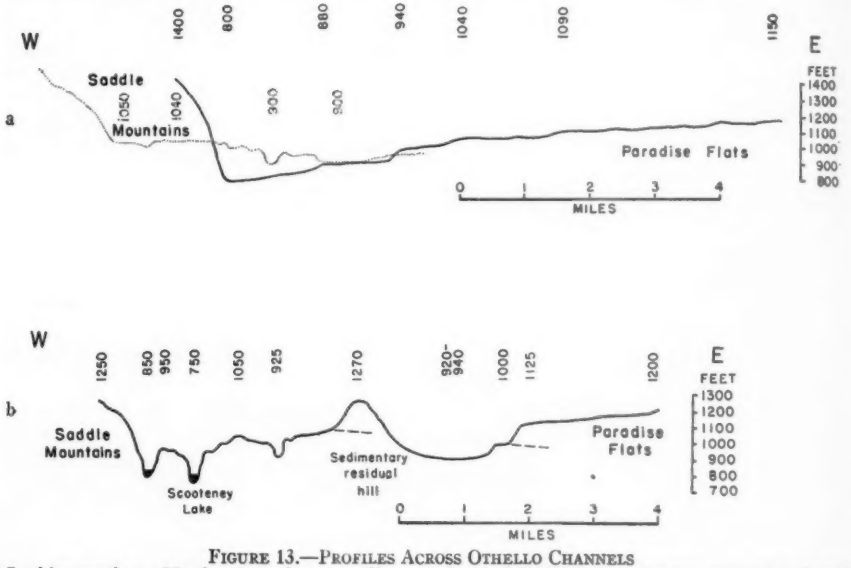


FIGURE 13.—PROFILES ACROSS OTHELLO CHANNELS

Looking north. a—Northeast-southwest profile across head of Channels, showing four successive channel floors. Taunton-Anson terrace profile, dotted, drawn for a parallel cross section 2¾ miles farther northwest. b—Nearly east-west profile where Channels transect Saddle Mountains anticline. About 7 miles southeast of section in Figure a.

(Scooteney Lake map), has a closure of 135 feet.[19] Like Drumheller, Othello Channels has a scarped residual hill of the once-overlying sedimentary rock, more than a mile across and its summit nearly 200 feet above the highest adjacent basalt scabland. East of this hill, Othello's largest channel is a mile wide and 100 feet deep in the sediments but is floored by basalt (Fig. 12).

East of Othello Channels is a plain a few miles wide determined by the flat-lying Ringold formation and constituting a prominent topographic feature from Warden 20 miles south to Connell (Pl. 1). The plain (Paradise Flats) overlooks glacial drainage channels from the Warden cataracts to Washtucna Coulee. The west-facing scarp is almost continuous from Lind Coulee to Esquatzel Coulee, and its base descends 200 feet in 20 miles.

The northernmost 8 miles of Othello Channels divergence (Othello quadrangle map) lies in the Othello basin where the record is largely in broad trenches without basalt scabland but with abundant evidence of the magnitude and transporting ability of the glacial discharge. Here four distinct, successive, terracelike channel-floor levels, separated by scarps, are recognizable in the field and on the Bureau of Reclamation contour map R4-5720, sheet 2 (Figs. 12, 13), each younger one trenched into the bottom of the immediately earlier one. All these channels shifted southwestward, in the plane of the cross section. The youngest, still intact and full of undrained depressions, has *no continuous gradient* of its bottom for its 8 miles of length. A little farther north, a remnant of the highest channel floor, the Taunton-Anson terrace, lies against the face of the mountain (secs.

[19] When the irrigation project is completed, this rock basin will be filled to its northern sill, 820 feet A.T. with a minor reservoir, and a channel in the southern and higher rim will be excavated 70 feet in maximum depth for 1200 feet through basalt to lead that water southward. Bottom water in the Othello glacial river here flowed 200 feet *uphill* for that distance, yet continued to pluck basalt from the channel bottom and carry it away.

10, 11, 14, 15, T. 15 N., R. 28 E.) and drops off 100–150 feet directly to the bottom of the lowest channel. This terrace is surfaced with at least 28 feet of gravel and boulders.

These channel floors are strewn with boulders. Excavators found basalt blocks so large that, even with modern equipment, the boulders could not be moved without blasting. They must have come from Drumheller, travelling across the sedimentary rock then flooring Othello basin for 10 miles on a gradient of about 15 feet to the mile. Not one of the channel terraces records a leisurely river of moderate size.

Another significant feature of this broad, headward part of the Othello divergence is a low, linear gravel mound a little above 1110 feet A.T. margining a part of Paradise Flats on the brink of its western scarp. A mile east of Othello town two active pits in it showed almost no boulders of basalt but many of caliche. The gravel is fore-set-bedded and dips dominantly southeastward diagonally away from the highest channel, at the top of whose bluff it lies. Immediately east of this deposit, the Flats have caliche in place to the grass roots. This gravel constitutes a channel-margin bar, a kind of natural levee, like the one southeast of Moses Lake town and is, insofar as known, the highest record of glacial water using Othello's scabland channels. The descent from the brink of Warden cataract to this place is 14 feet per mile. However, the surface gradient of the river probably was greater because the first altitude is on a channel floor, the second is a channel-margin record. Absence of basalt boulders suggests that the levee bar antedates excavation in basalt at Warden cataracts.

This complex of channels across the eastern tip of the Saddle Mountains anticline (Fig. 12), composed of three western, canyonlike, parallel slots cut in basalt and one mile-wide, equally deep eastern channel eroded largely in weak sedimentary rock (Scooteney Lake quadrangle map), poses a problem of origin and succession in development. Only if the preglacial surface of the sedimentary cover at the site of the Channels were lowest where the narrow notches developed would any such gashes in basalt be made. Only if glacial water demanded greater width than the 3 miles across the basalt portion or if these gashes were choked with ice would discovery of the soft-rock route occur. But by the time this happened the deep, slotlike canyons and rock basins were essentially complete. Downstream sills of two of them are but 10 and 30 feet higher than the basalt floor of the soft-rock channel and one is 50 feet lower.[30] No detailed succession of repeated floods through Othello is determinable at present, but probably the basin had no preglacial outlet and Lower Crab Creek glacial channel must date from the inception of Othello Channels.

In the lee of the isolated hill in the middle of the Othello group is a barlike mound, tapering to a point about 4 miles farther south (Fig. 12). It has a core of Ringold but is heavily veneered with scabland gravel. Its mapped surface has the same suggestive Λ's as are shown in the contours of the Bureau of Reclamation topographic map for the hills separating the three Quincy basin channels. They indicate subfluvial molding by a current crossing diagonally southwestward from overflow out of the eastern channel. After molding came the scarping and possibly 100 feet of channel deepening on each side to make the taper form. The deposit may be comparable in origin to the high gravel remnant in Bacon syncline.

Beyond the end of this barlike form, all Othello channels become one, the Ringold having been swept off the basalt for a width of 10 miles and a cover of gravel left, interrupted by a few minor scabland tracts. About 10 miles south of the isolated hill, this wide, stream-swept tract breaks up into separate channelways with flat-topped, steeply scarped, prow-pointed, isolated hills of Ringold separating them. Jackass Mountain (Eltopia quadrangle) is the southernmost of these residual elevations which divided the Othello Channels discharge, part going southwestward into the Koontz Coulee group and part southeastward into Esquatzel Coulee, but both eventually reaching Pasco basin (Pl. 1).

The Koontz Channels (Fig. 17; Pl. 1) consist of two groups, one with channel heads as much as 200 feet higher than, and truncated by the scarps of, those of the lower group. This truncation records at least two episodes of glacial-

[30] One must be on his guard in dealing with bottom gradient of these unconventional giant rivers. This 50-foot lower, downstream sill must be balanced against an upstream sill 50 feet higher than the soft-rock channel head.

river discharge out of Othello Channels and across the Ringold plain.

Yet these higher Koontz channels do not record the earliest glacial discharge across this plain. Streams that made no channels left a widespread mantle of boulderless caliche and basalt sand and gravel, containing granite, quartzite, and Ringold pebbles, and overlying caliche *in situ*, 50 feet or more above the floors of the highest Koontz channels. On the summit of White Bluffs, overlooking the Columbia 500 feet below (in T. 10 and 11 N., R. 28 and 29 E.), this glacial stream debris is fore-set-bedded, dipping up to 15° S. and SE. Its altitude, 900 feet A.T., is only 200 feet lower than the natural levee east of Othello town, nearly 25 miles upstream. The bedding and mantle character prove that berg transportation was not involved. Was this earliest record on the Ringold plain made by ordinary meltwater streams? Is it to be grouped with the high gravel of Bacon syncline, the unchanneled, boulderless sand and granule gravel in western Quincy basin, and the fine gravel apparently underlying the marked bars in Washtucna Coulee? Is it possibly preflood outwash?[21] Did it come from a completely filled Washtucna instead of Othello?

The featureless flat traversed by the Potholes East Canal (Fig. 17; Pl. 1) 2 miles west of Mesa (Scooteney Lake quadrangle) has a southward slope and, on the Bureau of Reclamation contour map, carries numerous Λ's of contour loops like those in Quincy basin. The flat is covered with about 10 feet of very coarse gravel containing many 1- to 2-foot boulders and locally overlying remnants of Ringold. This gravel came from Othello Channels and was carried at least 5 miles on a gradient of about 20 feet to the mile without the development of channels or bars. It seems that the near-by channels in basalt and Ringold did not then exist and that the flood water was too heavily laden to do any deepening on this gradient. Correlation of this flat with some of the higher, hanging Koontz channels seems feasible and would place it in the record of the second discharge out of Othello. If correct, there was no open Washtucna-Esquatzel Coulee at that time. Probable Ringold sediments beneath scabland gravel in lower Washtucna indicate that the plateau drainage system was well developed in basalt before Ringold deposition. This part of the pre-Ringold Washtucna-Esquatzel Coulee may have been still unexhumed when this second glacial discharge took place. Or it may have held a normal outwash of preflood age. Re-excavation came after deposition of the gravel cap on the flat west of Mesa and was effected by glacial flood water. Scabland channels were also initiated in later stages of this second flooding although most of them were deepened during later floods. Younger bars superposed on older bars in Esquatzel, described in later paragraphs, require at least a fourth flood.

The northernmost Koontz channel does not reach the Columbia. It opens westward on a Ringold plain at about 800 feet A.T. (Hanford quadrangle map) which descends to 700 feet 4 or 5 miles farther west and there terminates in the White Bluffs summit (Fig. 17) 300 feet above the river. This channel has a small distributary hanging about 75 feet higher at both ends; therefore it is not to be explained as a shallower contemporaneous spillway.

These channels differ from all others of the group in failing to reach the Columbia bluffs. Lack of channeling across the intervening plain indicates that a Columbia flood covered the flat when they functioned.

All other Koontz channels reach the bluff summits a few miles farther south and have hanging mouths. In Rs. 28 E. and 29 E. (Hanford and Scooteney Lake quadrangles) deeper channels truncate shallower ones to provide notches in the cliff summit ranging from 525 to 775 feet A.T., all within a 4-mile length of the bluff. Not all differences in altitude can be imputed to differing depths of water in contemporaneously operating channels and do not seem adequately explained by a lowering of the Columbia during one episode of Koontz channel activity. From the channel-mouth relations of the entire group, there appear to have been at least three episodes, each succeeding one finding a lower level of water (or debris) in the master valley.

[21] A Bureau of Reclamation test pit in SW ¼ Sec. 5, T. 11 N., R. 29 E., found, in descending order, 5 feet of sandy silt, 11.7 feet of coarse sand with caliche gravel, 2 feet of fine, sandy silt, 5.3 feet of basalt sand and caliche fragments, beneath which was at least 8 feet of very compact caliche.

Earlier Flooding in Quincy and Othello Basins

(Bretz and Neff)

For simplicity and clarity, discharge of Warden cataracts has been presented as part of the earliest glacial overflow from Quincy basin, coincident with that over the three western cataracts and with the discovery of basalt in any outlet channel bottoms. This was the writers' interpretation in 1952. Neff's subsequent studies, however, have uncovered evidence of earlier and higher floodings, the record apparently convincing but not detailed, and with some correlations uncertain.

Neff's evidence consists of hundreds of erratic blocks on the north slope of Frenchman Hills and the east slope of Babcock-Evergreen Ridge, some in clusters, some with associated gravel, up to 1360 feet A.T. In addition is a significant little stream-cut notch across the Evergreen summit a mile north of Frenchman Spring cataract, its floor at 1350 having a deposit of west-dipping fore-set gravel 5 feet thick and terminating at the western end in a cirquelike waterfall alcove, well shown on Sheet 2 of the U.S.B.R. map R4-5720.

A few boulders have been found at a comparable altitude along the east side of Quincy basin, and a tract several miles wide (east-west), denuded of its Palouse loess down to a caliche zone, margins the Willow Creek-Moses Lake trench from Rocky Coulee to Warden and lies below an altitude of about 1350. This flat has almost no southward gradient for the 15 miles of its length, and it disappears in the latitude of the cataract. It reappears at a lower altitude east of Othello town, south of the Frenchman Hills and Lind structures, with a caliche pavement under a thin loess, and there has stream gravel up to 1160 feet A.T.

These boulders and this flat in Quincy basin, 100 feet higher than the Warden cataract brink, indicate that the sedimentary cover at the site of Drumheller was sufficiently thick and extensive to have briefly held back the body of water responsible for flotation of the boulder-carrying bergs and the scour of loess from the flat. There must have been adequate volume to supply this discharge and the four spills across Babcock-Evergreen, three of which became the great cararacts in later floods.

Othello basin has some erratic boulders on the south slope of Frenchman Hills up to 1180 feet and apparently must have been sufficiently enclosed for their flotation. The residual hill in the middle of Othello Channel is good evidence that this basin's south rim originally stood high enough for such closure. A low summit notch on Frenchman Hills (Sec. 21, T. 17 N., R. 26 E. Smyrna map), altitude about 1370 feet A.T., separates boulders at 1360 on the north from boulders scarcely 2 miles distant on the south at 1180. A 10-foot greater depth of flood water in Quincy basin would have spilled across this preglacial divide. Since the caliche cap in the notch is intact and there is no scour on the slope down to 1180, water in the basin apparently never has stood higher than 1360.

Othello Channels has an accessory stream channel across Saddle Mountains ridge with a floor at 1150–1160 in a notch originally bottomed at 1190 or less (Fig. 12). A neighboring notch floored at 1210 was not used. The two lie about 3 miles west of the main Othello scabland (Scooteney Lake quadrangle map). Discharge through the channeled notch appears to have gone down a near-by, high-level Koontz channel in pre-Warden time.

At the time of this earliest-recorded Othello basin flooding with discharge across Saddle Mountains axis, either no well-opened Lower Crab Creek valley existed or the water gap of the Columbia farther west was closed. Ponding in Pasco basin, and therefore in the gap, cannot explain the discharge southward across the anticline. Possibly the great Taunton-Corfu slide was a factor, but it is much more probable that the course of Lower Crab was still largely unexcavated below the level of the Royal Slope, a broad structural terrace above 1150 and some 25 miles farther west. Initiation of both Lower Crab and Othello channels apparently came at this time.

Washtucna Coulee

Maps: Benge, Connell, Eltopia Haas, Washtucna quadrangle maps, U.S.G.S., R4-5720, sheet 3, U.S.B.R.

(Bretz and Smith)

Drainage

Glacial drainage from the north entered the preglacial Palouse River valley in mid-length

and followed it for about 25 miles, denuding more than 750 square miles of the Cheney-Palouse scabland tract. In the vicinity of Hooper and Washtucna, where the Palouse and Snake rivers were flowing parallel and about 10 miles apart, the distribution of scabland shows that a large portion of this glacial water escaped southward from the Palouse Valley across the divide and entered the canyon of Snake River, making scabland of the divide top (Pl. 1). Palouse River today is detoured from its former course along now streamless Washtucna Coulee into a striking, joint-determined scabland canyon which cuts through the preglacial divide. For some reason, perhaps incomplete removal of Ringold fill or of prescabland glacial outwash in the old course, Washtucna Coulee could not carry very much of the glacial water. Bretz insisted that the divide crossing was because of a volume far too great for Washtucna's capacity. Allison would doubtless explain the short-cut by an ice jam somewhere down the abandoned valley. Flint specifically referred the spillover to an aggradation of the lower Palouse Valley (Washtucna Coulee) until overflow was inevitable. He supposed "a very low route" available across the breached divide. Washtucna Coulee has another scabland short-cut to the Snake across the divide, 15 miles farther west—the streamless Devils Canyon.

Near Connell, about 35 miles down Washtucna Coulee, the character of the old valley changes considerably with entrance near that town of glacial drainage from Othello Channels. Thence downstream, it is known as Esquatzel Coulee. Some hitherto undescribed and some disputed features of Washtucna-Esquatzel are significant for a successful theory for the origin of scabland.

Staircase Rapids and its Bar

Staircase Rapids bar, 1–3 miles north of the town of Washtucna (Fig. 14; Pl. 1; Washtucna quadrangle map), lies in the mouth of a tributary entering from the north at the head of the abandoned portion of the prescabland Palouse river valley. The bar was made by a small, divergent strand of water escaping westward from the Cheney-Palouse tract and entering Washtucna along the short tributary to the old Palouse. The deposit, 150 feet thick, lies in the last 2 miles of this tributary but does not make contact with the east side. Flint (1938, p. 478–479) noted (1) south-dipping fore-sets in this bar as constituting "more or less flat-lying lenticular courses" instead of delta fore-sets, (2) terraces on the south face of the deposit which Bretz had described as a deltalike front, (3) cut-and-fill bedding, and (4) failure of fore-sets to parallel the eastern slope, also interpreted by Bretz as a constructional front. He concluded that the deposit is but a remnant of a more extensive valley fill, and held that the unfilled eastern part of the tributary valley mouth was the result of subsequent dissection because he found "a thin but distinct covering of identical material" on the eastern slope.

Allison (1941, p. 69) accepted Flint's interpretation although he upheld Bretz's view that the eastern front had "constructional lobings . . . ." because of "a reshaping . . . . [by] a later discharge".

Flint's interpretation of Staircase Rapids bar contains faulty reasoning and is in serious conflict with several lines of field evidence.

(1) If his general picture is to be internally consistent, his episode of aggradation must have begun, in this region, with deposition at the very bottom of the preglacial Palouse Valley and its equally deep major tributary, Cow Creek, which enters from the north. For the Staircase divergent spillover to occur from his "shallow" Cheney-Palouse river, the floor of Cow Creek valley had to be aggraded until it was close to the altitude of the divide at the head of the Staircase tributary (Fig. 14). At present that would require nearly 400 feet of filling in Cow Creek valley. This figure should be reduced by whatever deepening that valley had during scabland making. The three rival interpretations, however, consider present depths of the main preglacial valleys largely to date from prescabland time.

Only gradually could main valley bottoms have been raised by the procedure on which Flint insisted. Thus the mouth of the Staircase tributary gulch would inevitably have become back-filled from Washtucna's rising fill long before the Rapids could take origin. Four hundred feet of filling in Washtucna would have back-filled the tributary mouth 100 feet higher than the present top of the deposit and would

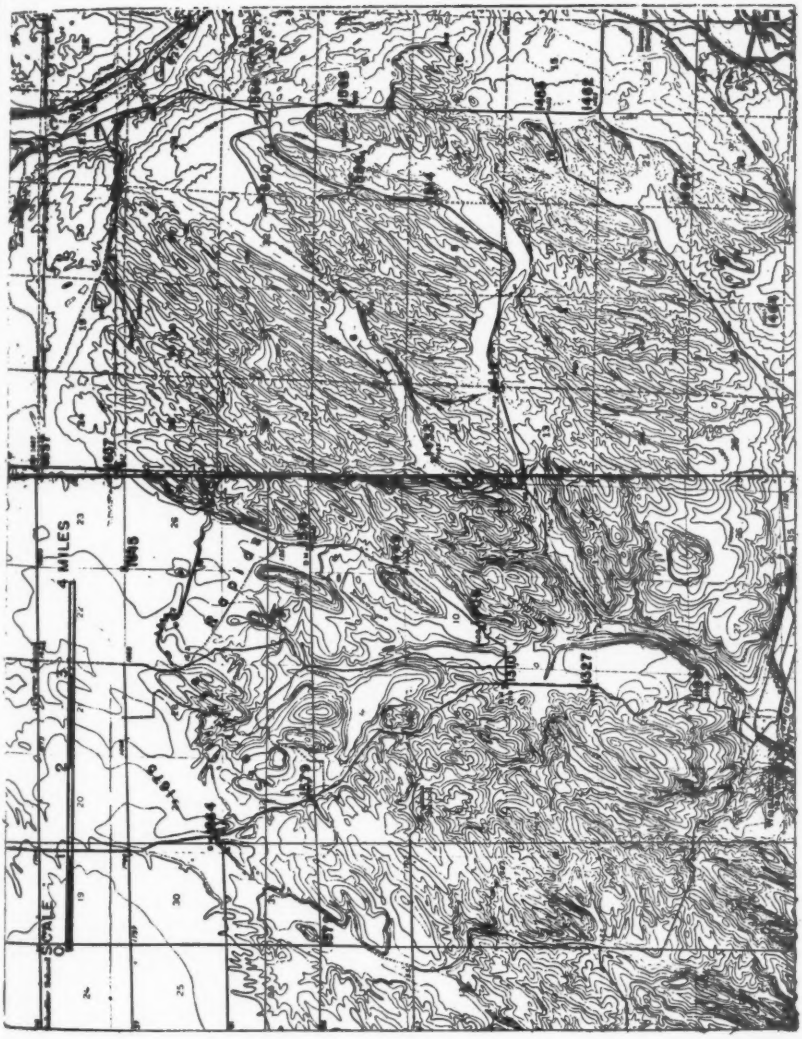


FIGURE 14.—STAIRCASE RAPIDS AND ENVIRONS

Part of Washtucna and Benge topographic maps, U. S. Geol. Survey. Contour intervals: Washtucna, 25 feet, Benge, 40 feet.

have produced only back-dipping fore-sets, if any. Only at the very last of Flint's episode of aggradation could any glacial water have been detoured from Cow Creek down Staircase. It would have found the little valley mouth already filled!

(2) The head of Staircase Rapids, interrupted by a small loessial island, is 3½ miles wide, and the brink of the uppermost falls is 1625–1650 feet A.T. The "bar" surface, 4 miles downstream, is 1300–1325 feet A.T. Flint took that surface and associated thin gravel to the east to be "remnants of a continuous fill graded to the fill in Washtucna Coulee". The gradient of any fill along the Staircase tributary valley therefore was 75 feet or more to the mile. It can be reduced to about 50 feet by considering the "bar" surface originally to have been 100 feet higher. Even this is an impossibly steep gradient. A proponent might conceive therefore that this channel never had a fill upstream from the "bar", that its scabland was made *during* the general aggradation, not afterward as Flint believed.

(3) It is impossible however to adjust this alternative to the fill-and-cut theory because a mile west of the head of the Rapids the western part of this divergent Cheney-Palouse strand entered its own separate spillway at close to 1675 feet A.T. (Fig. 14). For Flint's shallow glacial river ever to reach the head of this westernmost strand, a fill nearly 50 feet thick must have existed where the uppermost cataract brink now stands. Thus a once-continuous fill along the Staircase tributary to the "bar" is demanded, and its gradient must have been nearly 90 feet to the mile, instead of his estimated 13 feet for the aggradation in the Cheney-Palouse-Washtucna tract.

(4) There is, however, *no* gradient shown on the topographic map for the entire 2-mile length of the "bar". Instead, an undrained sag a mile long occupies the valley bottom at the *head* of the bar and, upstream from it, descent over the Rapids is 350 feet in a trifle more than 2 miles.

(5) Furthermore, the bar top is 100 feet lower than bases of loessial scarps 3 miles distant, almost directly across Washtucna Coulee. By Flint's view the flat-topped gravel remnant in this protected place should be as high as near-by scarp bases, 1400 feet A.T. From Flint's gradients the original surface of this deposit would have been 130 feet still higher.

(6) The only two scabland channels which cou'd have carried glacial water to erode the inner valley along the east side of the bar have heads but 25 feet lower than the low brink of the uppermost Staircase falls. Yet the deposit was trenched (?) 150 feet deep and not by Staircase discharge. Nor can the supposed trenching be ascribed to postglacial streams for, above the junction of the small channels with Staircase, the latter carries runoff from more than 16 square miles, the former from only 13. Eleven square miles of Staircase postglacial drainage enters the closed depression at the head of the bar and has not yet obliterated it.

(7) The two minor channels become one on the Benge-Washtucna quadrangle boundary, 1½ miles northeast of the bar. In each, a little above the junction, there is a flat-topped, undissected gravel flat (at 1400 and 1440 feet A.T.) with steep, downvalley-facing terminal slopes and fore-set beds descending about 100 feet to an empty channel floor. Gradient thence downstream is 75 feet in the 1½ miles, terminating in the silt-filled flat with intricate meanders locally dammed by one of the Staircase bar's lobations (Pl. 9, fig. 2). Gullying in this flat shows 18 feet of postglacial silt and sand with root casts. A large vertebral bone was also found. Upstream from the gravel flats are low but definite scabby basalt outcrops, one ledge constituting a waterfall more than 40 feet high with a plunge basin recognizable on the topographic map.

These features record the same kind and time of experience that is written in Staircase Rapids and completely disprove Flint's theory for the origin of the 150-foot trench alongside the Staircase bar.

(8) Finally, the volume and velocity of this minor divergence could never, by Flint's interpretation, have made the giant current ripples on the bar top (Pl. 9, fig. 2). The deeper part of the deposit could be a remnant of preflood, normal outwash (largely removed from Washtucna itself) and modified by later flood water, as Allison thought, but, if so, it must have been a backfill. The southward dip of it fore-sets does not agree with this view.

Allison's theory cannot explain the Staircase divergence unless the Cheney-Palouse tract was blocked with a huge ice jam in place of Flint's fill. For the westernmost distributary strand (item 3) to have operated, another ice jam must have formed along the steplike cascade, 3½ miles wide and nearly 50 feet thick at the upper end. The wonder is, here and elsewhere, that only the larger streamways became clogged; that minor ones carried off all the floating ice which entered them or were saved by timely failure of the major channels' blockades.

Other Deposits

The U. S. Geological Survey map (1954) of Haas quadrangle shows large patches of gravel (Qsu) associated with essentially bare basalt (Tcb) along both sides of Washtucna Coulee for 12 miles southwest of the coulee head. Near the head, gravel-covered areas extend a mile back on top of the preglacial coulee bluffs and touch the base of loessial scarps at 1400 feet A.T. The material is dominantly sand, granule gravel, and coarser basaltic debris, subangular to subrounded and stratified in thin lenticular foresets. It obviously is the marginal deposit of the highest flood, more than 100 feet above bluff summits and 400 feet above the coulee floor. Recognized in the map description as scabland gravel, it is as high as loessial scarp bases margining the Palouse-Snake divide scabland and therefore cannot record an early valley train in the coulee. A comparable gravel-covered area on this divide summit, 5–6 miles distant, south of Hooper (shown on Starbuck geologic map), reaches 1320 feet A.T.

Directly south of Washtucna town, a local hill of gravel rises from the coulee floor (about 1020 feet A.T.) to a summit more than 1200 feet A.T. (NE ¼ Sec. 33, T. 15 N., R. 36 E.) (Haas and Washtucna quadrangles). It lies in the lee (west) of scabland knobs on the south side of the coulee, but its summit stands free from the coulee wall except at the head. It is half a mile long, as is the fosse between it and the wall. The deposit terminates to the west in a rather abruptly tapering nose between the fosse and the coulee floor.

The Spokane, Portland, and Seattle Railroad has made a long cut 30 feet deep through the hill's lower, northern edge, and, although structure is no longer decipherable, composition is well shown. There is none of the size sorting on which Flint based his theory of aggradation by moderate streams. Boulders up to 8, 10, even 12 feet in diameter clearly are broken columns from the scabland knobs at the proximal end. No strata of boulders are indicated; the big pieces lying at random in the cobble and pebble gravel. At the distal end of the hill the material is all well-rounded pebble gravel, the finest material of the entire exposure.

This gravel hill, standing almost free from contact with the coulee wall and partially enclosing an unfilled tract back of it, was made by a scabland river flowing down Washtucna Coulee. It is not perched on the valley wall. Its summit is 75 feet lower than the lowest spillover across the Palouse-Snake divide. The fosse is in no sense a gully of postscabland origin. The hill is a bar of coarse detritus in the protected lee of knobs which contributed boulders to it only under the urge of a stream adequate to detach and transport them. That stream was at least 200 feet deep.

At Sperry (or McAdam) siding (Pl. 1; Pl. 12, fig. 2; Haas quadrangle) a deposit of fine-textured, well-sorted gravel, with a few strata of nut- to egg-sized pebbles, masks the north wall of Washtucna Coulee. Texturally and topographically it would fit the fill-and-cut theory, but its structure is that of long foresets dipping 20° throughout the entire 40-foot pit wall directly *into* the coulee *from* the higher scabland ledges. The dip parallels the slope but is a little steeper (*see* Flint, Pl. 4, figs. 1, 2). The fore-sets, reaching to the bottom of the coulee, prove that Washtucna had no fill (above present bottom level) when these foresets were made. The valleyward slope is essentially constructional. It may have a core of older gravel.

There are two gravel deposits near the east end of Washtucna Lake (Pl. 1; Haas quadrangle). One, at Harder siding, is about a mile long. It lies against and largely covers the north wall of the coulee. No section was afforded, but the form is significant. The deposit has a flattish top which breaks smoothly over into a steep valleyward slope. But that top is

not level! Along its length, it slopes down to the west, and the departure from horizontality is 3° (Pl. 12, fig. 3). In other words, its east end lies approximately at the top of the coulee wall, and its west end, a mile distant, not far above the bottom. Should such a form be called a terrace?

The other deposit near the lake lies south of Wacota Siding in the lee of scabland eminences. Its surface slopes even more steeply westward; the steepness increasing at the terminus which reaches lake level at high-water stages. A profusion of big boulders lies on the floor of the pit in this deposit, but no structure was seen. The form and size are perfect for subfluvial shaping of flood debris lodged in this place. The location down in the basin of this lake without outlet is an awkward fact to fit into Flint's theory.

About midway between Kahlotus and Estes (Connell quadrangle map) a large gravel deposit conceals the northern scabland wall for about a mile. Three pits in the base of its slope yield a well-sorted, pea-size gravel. The bedding at the west end of the deposit dips 20° toward the valley through 40 feet of exposure. In a few places interrupting short fore-sets dip away from the valley. Two other pits show long fore-sets with a 5-degree dip toward the valley and a 5-degree dip along the valley, respectively. The top of this deposit, above which is scabland, is flattish and descends westward with a 3-degree slope throughout its length and tapers out for no apparent local reason. Should this form and structure be interpreted as simply a terraced remnant of a once-continuous fill?

Just west of Sulfur Lake[22] in Washtucna Coulee (Pl. 1; Connell quadrangle map), a smoothly rounded, east-west elongated, gravel hill about half a mile long stands out on the coulee floor. Its top is more than 50 feet above the adjacent coulee bottom, more than 100 feet above near-by Sulfur Lake, and more than 30 feet above an undrained depression (a fosse) on the north side of it. There are no exposures. Most of this nearly mid-channel hill is lower than the site of Connell; in other words, it stands *down* in the long, closed depression of Washtucna Coulee. It is not "perched" on the valley slopes. Bars are equilibrium forms in stream beds. Some may be of erosional origin, shaped from older alluvial deposits. But only a stream capable of making the 12-mile long, 100-foot deep depression could have sculptured this hill from a former continuous valley fill.

A fill in the mouth of Hardesty Coulee (Pl. 1), which enters Washtucna from the north just east of this mid-channel hill, has a flat, level top close to 1100 feet A.T. and rests on basalt in the gulch at about 950 feet. Much of it is granule gravel of bluish-black basalt, but it has strata of nut- and egg-size pebbles. There are cobbles and small boulders of quartzite and granite on the coulee floor. Sections are poor and show only indistinct horizontal bedding. Buried beneath portions of this gravel are silts and sands, the basal part containing a thin conglomerate of basalt and the whole taken to be a remnant of the Ringold formation. If so, Hardesty Coulee was a valley eroded in basalt essentially to present depths when the Ringold was deposited.

Although there is no definite knickpoint along lower Hardesty, there is a pronounced narrowing where the stream enters and crosses this gravel fill, a narrowness that has persisted while and since, according to the fill-and-cut theory, Washtucna Coulee was essentially cleaned out by the subsequent erosion of a distributary from the Cheney-Palouse tract. Trenching of the gravel deposit has contributed to filling of the Sulfur Lake depression which nevertheless is still 100 feet deep. Debris from Hardesty cannot get out of that hole.

The fill at the mouth of Hardesty has a good terrace form 150 feet above the level of the Connell "blockade" although more than 100 feet below the base of loessial scarps. Its composition indicates consanguinity with those other "terraces" whose flat tops slope down-coulee. Its constriction of the tributary valley seems quite out of harmony with the picture of essentially complete valley re-excavation of Washtucna by a river of moderate size.

Rattlesnake Coulee (Pl. 1; Connell quad-

[22] A serious flaw in Flint's explanation, noted by Allison, is its failure to account for the Sulfur Lake depression. The basin is 12 or 13 miles long and more than 100 feet deep in the bottom of one of the valleys whose fill he believed was removed by later normal glacial river work. It has bare scabland buttes in mid-channel at the lake.

rangle map) enters Washtucna about 2½ miles east of Connell, and its discharge thence flows east (upvalley) for 4 miles to enter Sulfur Lake. Its mouth contains a gravel deposit made back in it from the larger valley. A pit here shows bedding fore-set up the tributary out of the main valley for the full height of the section, 20 feet. No horizontal members were seen. Some of the fore-sets are at the angle of stability, others a little less steep. A seam of light-colored sand conforms perfectly with fore-set gravel above and below. Under the bluish-black, fore-set, fine gravel is a foot or so of older, stained and softened gravel and sand with horizontal bedding, taken to be a basal part of the Ringold. On both sides of the coulee mouth, the top of this deposit is a rounded form with a depression, now drained, up to 30 feet deep between it and the scabland wall to the north.

Esquatzel Coulee

Maps: Connell, Eltopia, Scooteney Lake quadrangles, U.S.G.S.: R4-5720, sheet 3, U.S.B.R.

(Bretz, Smith, and Neff)

Providence Coulee, entering Washtucna-Esquatzel at Connell (Pl. 1), is shown even by the 50-foot contour interval of the Connell 1 to 125,000 topographic map to have a pronounced narrowing below 1000 feet A.T. Providence Coulee crosses three formations. The valley is maturely open on the loess-covered higher basalt country; becomes a ravine across about 2 miles of Ringold (Paradise Flats topography); and, at its entrance into Washtucna, narrows to a trench across a back fill of gravel occupying the ravine mouth. Long uninterrupted fore-sets dipping diagonally up the coulee for the full height, about 40 feet, of a section along the Connell Branch of the Northern Pacific Railroad are clearly visible from the highway a mile distant. Neither these nor any others of the long, unbroken fore-sets listed above can be explained as parts of a gradually growing valley-train deposit. Can they reasonably be interpreted as erosonial remnants of deltas?[23]

[23] Oral suggestion from James Gilluly in 1927, as an attempt to escape from the bar hypothesis.

Pits in this deposit show that the long fore-sets unconformably overlie finer gravel in which ordinary cross-bedding occurs in nearly horizontal courses about 2 feet thick. These courses dip about 5° NE. They may record earlier, nonflood deposition. But it is difficult to explain this trench if the gravel is pre-flood unless the almost complete cleaning out of Washtucna is ascribed to flood erosion.

Connell stands on a fill of fine gravel which constitutes a divide in the old river valley (Connell quadrangle map). Drainage from Providence Coulee goes southwest down Esquatzel but run-off at Connell also goes eastward *up* Washtucna, descending more than a hundred feet in 7 miles to reach alkaline Sulfur Lake. It is 180 feet to basalt at Connell. However, the altitude of this mid-coulee divide, about 850 feet A.T., is far too low to allow consideration of this fill as the blockade which caused the Devils Canyon short-cut to the Snake beginning at 1250 feet, or the Palouse Canyon transection whose highest scabland is about 1400 feet. According to Flint's theory, the Connell fill can be only an erosional remnant of a valley deposit which originally must have reached to the highest scabland at that town, about 1100 feet A.T. A rising gradient of such a fill up-valley would account for the two canyoned divide crossings, Devils at 1250 feet and Palouse at 1400 feet.

Part of Connell is built on a low, rounded, gravel mound in mid-valley, composed largely of granule gravel and certainly not the deposit of a vigorous stream. Evidence of mid-valley, gentle-current deposition has been consistently sought. Lateral deposits in protected places were unacceptable. This site of Connell comes nearest to answering the demand. The transporting current, however, may not have come down Washtucna. Contour maps by the U. S. Geological Survey and the U. S. Bureau of Reclamation show a flattish tract which descends eastward from the Othello divergence that enters the mouth of Washtucna and the head of Esquatzel. It begins about 5 miles west of Connell at 950 feet A.T. and descends 100 feet in reaching that town. Continued farther east up Washtucna, the slope descends another 100 feet in 7 miles to Sulfur Lake. Lack of adequate exposures along this 12-mile slope

renders an interpretation speculative. But the nearly continuous and nearly uniform slope and the character of the deposit in pits at Connell suggest that it is a record of backfilling from Othello discharge subsequent to Washtucna's glacial river. This would help explain the Sulfur Lake depression. Silts overlying coarse scabland gravel at the lake may be contemporaneous with deposition of the barrier or may record a postglacial lake.

Supporting the view that the gravel at Connell and in the mouth of Providence Coulee came from the west (from Othello Channels) are strong eastward components in the long fore-sets dipping back up Providence and the occurrence of other eastward-dipping fore-sets in a small pit not far west of the railroad section.

One serious objection to this interpretation is that the Connell deposit is more like a valley bottom fill than a delta. Another is that the bedding in a shallow pit at Connell is fore-set westward. It is conceivable, however, that the divergent strand from Othello Channels was succeeded by renewed functioning of a glacial Washtucna which reworked the upper part of the coulee bottom deposit, giving it the aberrant slope and a superficial fore-set structure with westward dip.[24]

Unlike Washtucna, Esquatzel Coulee varies greatly in width, ranging from an eighth to three quarters of a mile between walls. The towns of Mesa (Scooteney Lake quadrangle) and Eltopia (Eltopia quadrangle) mark two narrow places, and there are three others less striking. Their significance lies in the massive gravel deposits constituting very local hills which constrict the old valley in these narrow places. They are not "perched" deposits. This difference between Esquatzel and Washtucna is a direct consequence of occupation, at least from Mesa southward, by one of the divergent strands supplied from Othello Channels.

Two of these gravel hills overlapping the coulee walls, one a mile south of Mesa, the

[24] A gravel deposit along the highway 4 miles directly south of Connell is only tentatively interpreted by Bryan (1927, p. 27), Flint (1938, p. 518), and the writers. It occurs on the slopes of Old Maid Coulee, a tributary entering Esquatzel from the east. Its altitude, 1050 feet A.T., is about that of the highest scabland in the larger valley. Flint reported a B soil horizon "consisting of 3 to 5 feet of calichified silt and firm pure caliche", underlain by a dominantly basaltic gravel ranging from granules up to boulders 3 feet in diameter, and overlain unconformably by Touchet silts. Quartzite and caliche pebbles and sand with feldspars, micas, *etc.*, occur in the considerably decayed basalt gravel. Fore-set bedding dips southwestward, down along Old Maid Coulee.

But there is no scabland in this coulee nor any scarp-bounded notch at the head where glacial water might have overflowed from Washtucna Coulee and introduced the foreign material. The caliche cap also sets the deposit apart from all scabland gravel ever described. Nevertheless, it probably records a meltwater stream across the plateau. Its 3-foot boulders indicate vigorous erosion somewhere enroute, and the caliche pebbles suggest a stream that overran surfaces above valley bottoms.

PLATE 13.—SCABLAND DETAILS

FIGURE 1.—Shoulder bar. Note parallelism of ripple marks with growing (northern) front of bar and their extension from summit down southern and western slopes to constitute the crenulations shown on Haas topographic map (*see* Fig. 21). Production and Marketing Adm. photo CCH-16-197.

FIGURE 2.—Part of bench between Babcock Ridge and Columbia River. Three contrasted surfaces are shown: (1) the northern part without scabland, (2) older scabland, and (3) younger scabland. TT—Trinidad or Crater terrace. WSD—Willow Springs Draw. G—gravel pit. B—gravel bar associated with older scabland. O—gravel marginal to younger scabland. Northward extension of the lower cliff up into the mouth of Willow Creek Draw shows that Columbia Valley and the Draw had present depths before any scabland was made on the bench. U. S. Army aerial photograph.

PLATE 14.—BARS NEAR MESA

FIGURE 1.—Interrupted panorama of terminal portion of bar south of town. Looking north. Fore-set bedding for full height of each pit dips as shown by arrows. Only slight gullying has occurred between bar on left and older gravel on right. Photo by H. T. U. Smith.

FIGURE 2.—Face of bar north of town. Looking west and showing the gulch across it, the southward slope of the summit, and the down-coulee tapering out of the deposit. Photo by H. T. U. Smith.

FIGURE 3.—Highway section through bar north of town. Looking south. Structure shows that valleyward slope can only be constructional. Photo by H. T. U. Smith.

FIGURE 1

FIGURE 2

SCABLAND DETAILS

FIGURE 1

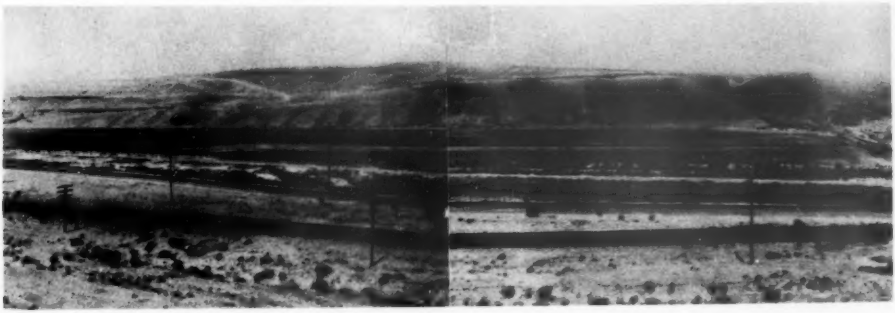


FIGURE 2

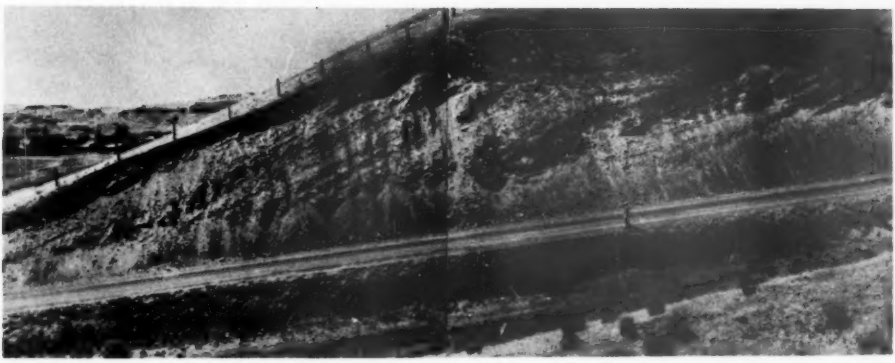


FIGURE 3

BARS NEAR MESA

FIGURE 1

FIGURE 2

FIGURE 3

FIGURE 4

GRAVEL BARS SOUTH OF SNAKE RIVER

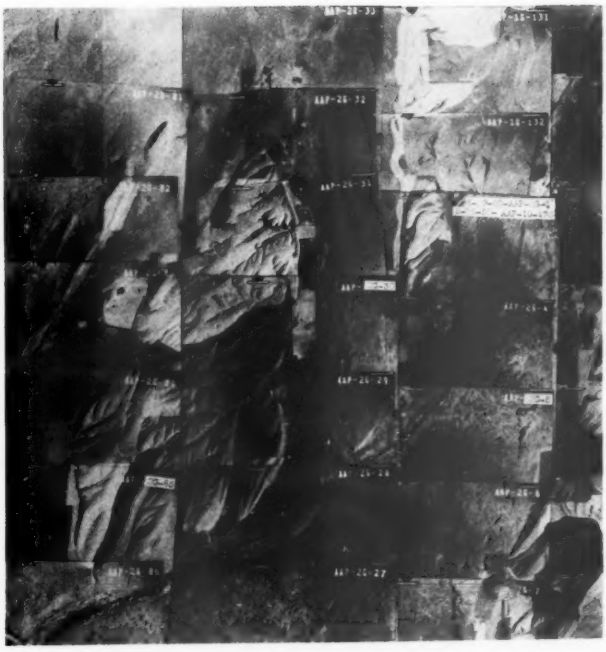


FIGURE 1

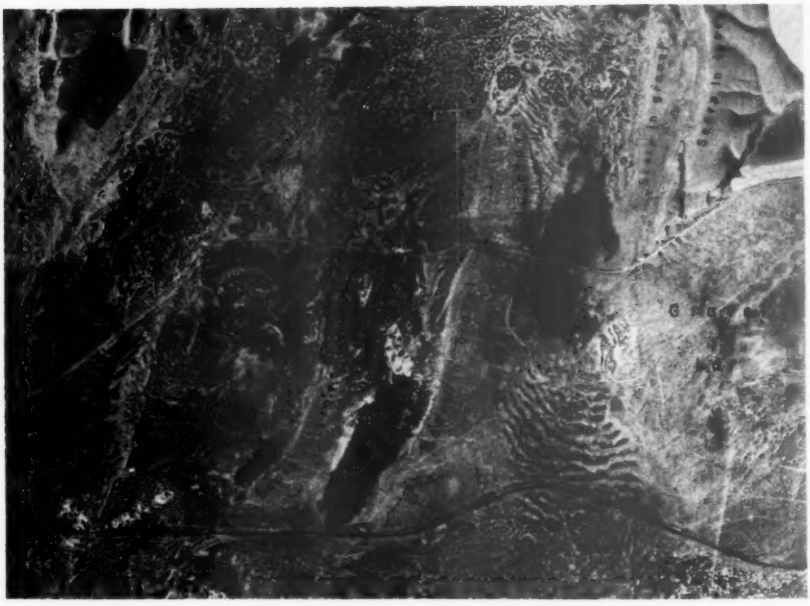


FIGURE 2

SCABLAND FEATURES IN MID-WIDTH AND MID-LENGTH OF CHENEY-PALOUSE TRACT

other a mile north, have excellent sections (Pl. 14). The southern hill is 50 feet high at its blunt, downvalley terminus where it stands free from, but parallel to, the valley wall. On the valleyward side, the hill has three or four lateral lobations cut by the highway and one by the railroad. Even better sections are afforded by two large pits.

All these cuts show long, uninterrupted, deltalike fore-sets dipping downvalley. The larger pit wall (Pl. 14, fig. 1) shows 40 vertical feet of such beds dipping about 20° and striking essentially tangent to the curved end of the hill. There are no horizontal strata and no granule gravel, sand, or silt members. The gravel is all moderately coarse and moderately worn basalt, with some cobbles and a few boulders.

The form of the deposit is equally enlightening. The hill's elongated, rounded summit parallels the valley wall, a sag or fosse separating them. The unterraced valley slope against which the hill rests is of gravel, or at least is covered with gravel. It is marked by broad shallow gullies, features which the hill itself does not possess. Exposures in this slightly gullied valley slope show a fine-textured gravel with but few medium to large pebbles. The fine gravel is apparently older than the coarse. Yet there are definite constructional moundings on its slightly gullied surface, and both deposits are judged to be bars.

The deposit north of Mesa has a gulch cut through it (Fig. 15; Pl. 14, fig. 2). It is a hill of gravel built against higher, gravel-covered slopes, a fosse with its central part still undrained lying between. The hill summit

PLATE 15.—GRAVEL BARS SOUTH OF SNAKE RIVER

Photos by H. T. U. Smith

FIGURE 1.—Pit section in northwest part of shoulder bar. False and true fore-set bedding in lower left. One bevels the other. Note a fence (F) in Figures 1, 2, and 3.

FIGURE 2.—Pit section about half way down bar front

FIGURE 3.—Pit section at toe of bar

FIGURE 4.—Trickle bar south of Lyons Ferry

PLATE 16.—SCABLAND FEATURES IN MID-WIDTH AND MID-LENGTH OF CHENEY-PALOUSE TRACT

FIGURE 1.—Scabland with loessial islands. Scarified but not deeply channeled scabland with two scarped and prow-pointed loessial islands carrying unmodified minor preglacial drainage lines above flood limits. One of them also has a minor channel across it. Light areas are wheat stubble, dark areas are summer fallow or fields newly plowed. Part of T. 16 and 17 N., R. 37 E. From Production and Marketing Adm. mosaic 7, Adams Co., Wash.

FIGURE 2.—Giant current ripple marks. The local scabland "channel" is 1½ miles wide between two small loessial islands, part of each showing. A larger "channel," containing these islands, is 8–9 miles wide and is bounded by two larger islands. Overall width of the Cheney-Palouse tract at this place, downstream from all distributary routes except Washtucna Coulee, is about 15 miles. Figure 1 shows the general character of island and scabland relations a few miles farther south.

Width of the small "channel" of the particular "stream" which left the concentrically curved current ripples is a quarter of a mile, approximately that of Snake River in flood today. It appears to be the product of a later and smaller flood which here traversed and incised an earlier gravel deposit. On the right is a double scarp, the upper part in loess, the lower in the older gravel. This ripple-marked "channel" is only part of an anastomosis, the gravel scarp and ripples of a contemporary channel shown on the left. Relief of the ensemble demanded three curves in the railroad to reduce grades, two of them shown in the photo. The higher gravel deposit on the right is thus shown to have a broadly rounded terminus of bar, not terrace, character in the lee of the loessial island. Profiles supplied by the C.M. and St. P.R.R., indicate that the railroad grade rises 20 feet in crossing from the west side of the area to round the tip of the bar to the east. Scaled from the photograph, the wave length of these ripples is 200 feet. Approximately Sec. 18, T. 18 N., R. 38 E.

Production and Marketing Adm. photo AAP-1G-125.

undulates through minor sags and swells that fall into a flattish profile when viewed from an equal height on the opposite side of the coulee. and strikes parallel to the frontal slope. Cobbles and boulders are not abundant, but pebbles are large almost throughout. A pit at the south, lee

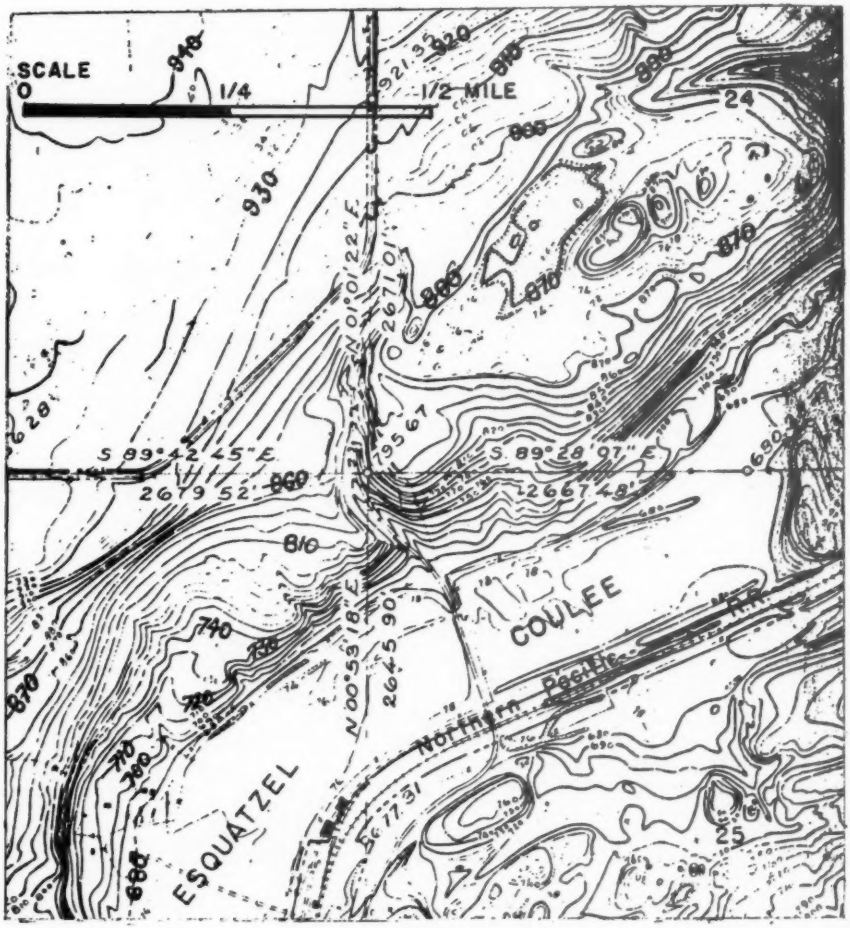


FIGURE 15.—BAR NORTH OF MESA

Showing the flattish free terminus (contours 700 to 740) and the undrained fosse (870 contour). Part of U. S. Bureau of Reclamation map G-5473, contour interval 2 feet.

This profile descends southward, downvalley, with a slope of 4°. The hill narrows and lowers to its terminus, free from the valley slope. Overall length is 3000 feet.

The road along the gulch affords four good sections, all of which show delta-type fore-sets from top to bottom (Pl. 14, fig. 3) with dips prevailingly of 20° toward the valley bottom

end of this valleyside hill shows rather fine gravel, with sporadic small boulders. It also shows a lens of silt and sand 3 feet thick, lying horizontally in the gravel. This could well be a deposit in a quiet water hole in the lee of the bar during its growth. A few days at the most would suffice for its settling.

It is argued here that the fore-set beds do not

record addition of their gravel *down* the slope from *above*. The contributing current was flowing horizontally past the growing bar face; the gravel was travelling more largely *along* the slope than *down* it.

The irregularly mounded valley slope back of and above the fosse has broad, shallow gullies. The hill, which is interpreted as a younger deposit, has only a few narrow short gullies, other than the transecting gulch.

Neither of the Mesa bars has a scabland channel leading to it. The erosional pattern of the adjacent upland in both cases is elongated subparallel to Esquatzel and the two bars.

Just south of Mesa, Esquatzel Coulee breaks up into an anastomosing pattern of channels which at Eltopia, 8 miles farther downstream, is 5 miles wide. Jackass Mountain, an isolated Ringold hill in this complex, stands 200 feet above closed, channel-bottom depressions on both sides. The course of the preglacial Esquatzel margins the eastern side of the complex (Fig. 17). Most of these interconnected channels converge a little south of Eltopia, but here another break-up of Esquatzel makes two dominant, diverging channels, almost equally deep and separated by a broad hill of Ringold 5 miles long and 200 feet above its bounding channels. The eastern channel, deeply filled with sand and gravel (U.S.B.R. borings), apparently records the preglacial Palouse. Beyond this hill channels disappear in the blowsand of the Pasco lowland.

This complicated glacial-river channeling, well shown by 10-foot contours on the U. S. Bureau of Reclamation map R4-5720, sheet 3 (*see also* Eltopia and Wallula quadrangle maps, U. S. Geol. Survey), extends from 750 feet A.T. near Mesa down to an altitude of 500 feet in a distance of 17 miles; all of it is well below the high level of Lake Lewis as defined by Allison and Flint.

The two Mesa bars stand down on the main channel floor. Water in Pasco basin apparently did not then, nor later, reach the 1100-foot scabland channels at the summit of the Wallula watergap cliffs, 9 miles distant. Yet the complicated channeling was done by a very large glacial river after the Touchet silts were deposited, and probably when both Othello and Washtucna were contributing to Esquatzel. If Washtucna glacial river was then flowing, Upper Grand Coulee had not yet been completed or was temporarily blocked by the Okanogan lobe.

These channels clearly record a post-Touchet flooding down Esquatzel, and, if Allison is correct in making Touchet silts younger than the Snake Canyon gravel bars, Touchet (*i.e.* Lake Lewis) sedimentation, although glacial, must be interflood in age.

Pasco Basin

Maps: Coyote Rapids, Eltopia, Hanford, Pasco, Pricst Rapids, Prosser and Wallula quadrangles; U.S.G.S.; and R4-5720, sheet 2, U.S.B.R.

(Bretz, Smith, and Neff)

All plateau scabland drainage, entering the Columbia to the west and the Snake to the south, converged to the Pasco basin, the largest and deepest of the structural-topographic basins of the region (Pl. 1). Nearly 500 square miles of it lies below the 500-foot contour. The Columbia River at its outlet, the Wallula Narrows across the Horse Heaven Hills anticline, is 300 feet A.T. Scabland extends up on the walls of this narrows to 1100 feet A.T. or more, but the basin upstream is largely empty.

Flint believed (1938, p. 517) the high Wallula scabland was made because the Pasco basin was filled with debris up to "at least 900 feet". This fill was of proglacial origin but was earlier than the rise of Lake Lewis, *i.e.*, earlier than the valley-filling, loess-scarping, scabland-making episode. It must have caused aggradation in tributary valleys back on the plateau, but about this Flint gave no details. However, most of this early basin and valley fill, its cause unspecified, was removed before the widely distributed lacustrine Touchet silts (contemporaneous with the plateau gravel) were deposited in Lake Lewis, therefore before plateau scabland erosion occurred.

The high Wallula scabland, although "well developed", is thus much older than any other in the entire region.

Allison (1941) asserted that the subfluvially molded gravel deposits associated with

scabland in Snake River canyon had been derived from "high-terrace remnants" of a single, great, prescabland terrace system which he correlated with "terraces" in the Pasco lowland and, with declining elevations, farther down the Columbia. He visualized this system as filling the basin up to 700 or 750 feet A.T. But he was unsure of the time of dissection. It could have occurred by "ordinary erosion . . . between two distinct glacial stages" or it could have happened "during a single stage of glaciation before the Lake Lewis-ice-jam stage was reached" (p. 71). "If Lake Lewis and the Touchet beds are later than much of the gravel, then the rise and decline of Lake Lewis cannot be the cause of the deposition-excavation sequence in Flint's fill hypothesis. Possibly Lake Lewis or equivalent appeared twice . . ." (p. 66), but Allison could find no evidence to support this idea. The cause of the earlier alluviation is not stated. An ice jam in Wallula Narrows is suggested for the (later) 1100-foot scabland and 900-foot gravel deposits at Wallula Narrows.

Because the Atomic Energy Commission had done much construction in the northern part of the Pasco lowland since the last field study there, permission to enter the restricted Hanford area and examine the open pits was secured prior to the present study. Because Bretz's earlier descriptions of forms and relations had not carried conviction to some geologists that the gravel hills there were giant river bars, it was hoped that perhaps composition and structures shown in the A.E.C. excavations would do so.[25] Most of the excavations seen were shallow and obscured, but data from them support Bretz's idea of bar origin. A pit in a high gravel tail to Gable Mountain (SE ¼ Sec. 28, T. 13 N., R. 27 E., Hanford quadrangle) showed that the flattish-topped deposit is composed of coarse sand of mixed basaltic and granitic debris, horizontally bedded, very evenly stratified, and carrying a cover of cobbly and bouldery gravel. It is an erosional remnant extending from about 650 feet A.T. down at least to 600 feet, its streamlined shape being the result of the glacial Columbia's erosion when the mountain was an island in the river that made the associated great bars. It is presumably of prescabland age and may well be correlative with the valley fill of granitic sand at and upstream from Crater. The coarse overlay on its summit and slopes is a derivative from the scabland of the mountain, acquired when the deep channels (400 and 425 feet A.T.) on either side of Gable Mountain were made and this deposit of fine gravel shaped. In this sense, it is a terrace, eroded to bar form and armored with scabland gravel. The Columbia River surface just above Coyote Rapids, 4–5 miles upstream from these empty channels (*see* Bretz 1927b), is 400 feet A.T., and the streamlined gravel hills—*i.e.*, bars—which shut them off from the river are 100 feet higher. A spring in the bottom of the deepest depression (400 feet A.T.) fluctuates with stages of the Columbia River.

The "terrace" in the western part of the Coyote Rapids quadrangle[26] is shaped like a great spit projecting southeastward across ranges 25 and 26 E. into the lowland from a high basalt anticline (Umtanum Ridge) on the Priest Rapids quadrangle (Figs. 16, 17). At the west end, this "terrace" has steep scarps descending 200 feet toward the north and 125 feet toward the south. Its broad summit is 800 feet A.T. here but descends along its length about 100 feet, and its scarps become more gently sloped toward the southeast. In mid-length it is cut across by a channel 50–75 feet deep. Its length, 13 miles, projects more than half way across the Columbia valley bottom, separating the river on the north and northeast from about 50 square miles of lowland known as Cold Creek Valley to the south.

Three pits and one road cut afforded sections in this "terrace". Two pits showed well-worn, well-sorted pebble and cobble gravel, 50 per cent bluish-black basalt, with a few boulders. The bedding is horizontal without even shallow fore-sets. The highest of the three pits (approximately in NE ¼ Sec. 34, T. 13 N., R. 25 E.) on the brink of the 200-foot northward drop-off toward the Columbia has

[25] No maps were allowed in the restricted area, and the road system had been so greatly altered since the Commission took over that many locations could be only approximated. In addition, many roads were closed to private cars.

[26] Used by Allison in reconstructing his Pasco basin fill.

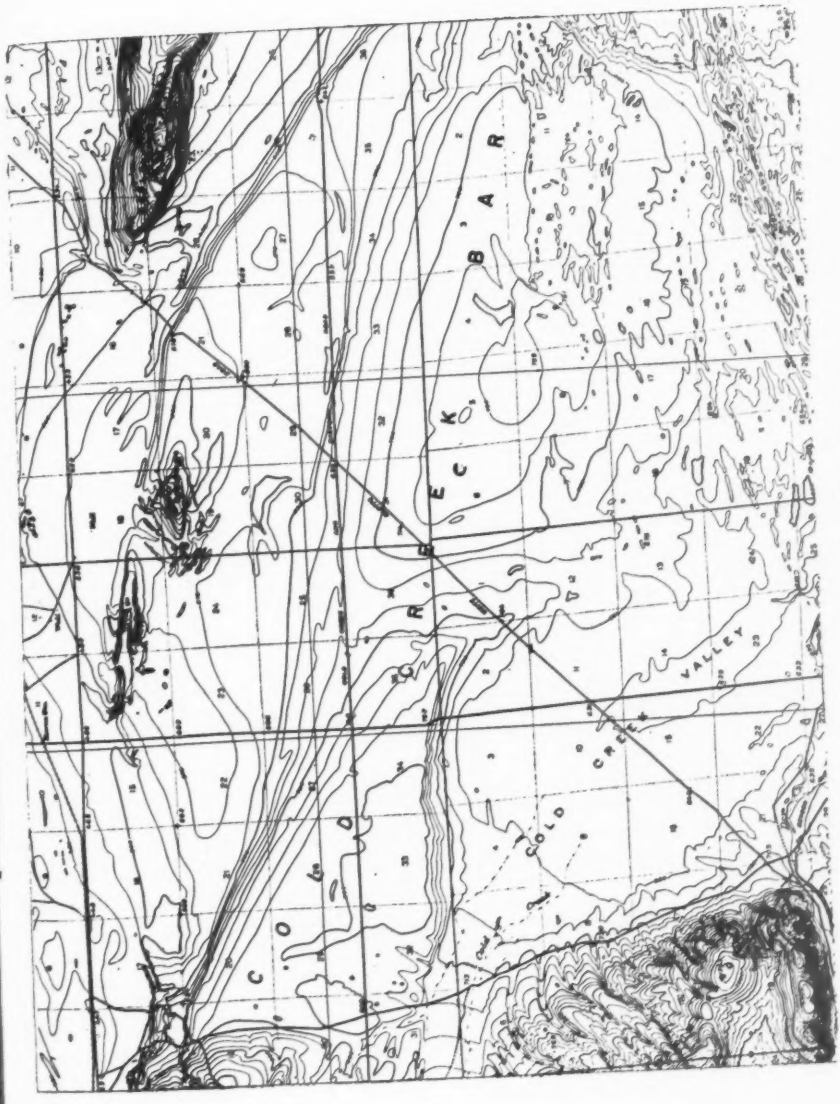


FIGURE 16.—COLD CREEK BAR
From Coyote Rapids topographic map, U. S. Geological Survey. Contour interval 25 feet.

only long deltalike fore-sets dipping southward, away from the river and into the deposit. Much of its material is composed of fine-textured, dirty gravel alternating with clean, openwork cobble gravel. Boulders of basalt and granite 3 feet in diameter and one basalt boulder 8 feet in diameter indicate transportation by floating ice. Because the freshness of the material also indicates a scabland derivation, this "terrace" was added to, if not wholly built, during scabland history.

The lee slope could be examined in section in only one place, the steep drop-off away from the Columbia at the southeast terminus, approximately in NW ¼ Sec. 13 T. 12 N., R. 26 E. The long road cut here showed only loose, dark, coarse sand with few cobbles and boulders, material typical of lee slopes of scabland bars. No structure showed to check that interpretation.

The character of "Cold Creek Valley" on this Coyote Rapids quadrangle must not be neglected in understanding the origin of the "terrace". The creek, draining about 50 square miles of the high, arid basalt country to the west, has built a large fan where it enters the "valley", a fan that spreads out abruptly from a width of half a mile at the head to eight times that width in a distance of 3 miles; the creek becoming lost in its fill. These proportions (Figs. 16, 17) show that the "valley" is only a part of the lowland, a tract partially enclosed by the giant, spit-like form. It remains for objectors to show that Cold Creek Valley, the 50-square-mile area lying in the lee of this "terrace" (Coyote Rapids, Hanford, Pasco, and Prosser quadrangles), can possibly be an erosional product of Cold Creek, as it must be if the "terrace" is but a remnant of a valley-bottom fill. It is significant that the great collection of iceberg mounds of glacial till on Iowa Flat (Bretz, 1930b, p. 410; Allison, 1941, p. 673) is less than 10 miles farther south, lying in the protected lee of this spitlike deposit and extending from 625 to 825 feet A.T. Cold Creek Valley was in existence when these bergs stranded at altitudes representing the full vertical range of the "terrace".

This extraordinary display of hundreds of till mounds, up to 20 feet high, and 100 feet or so in diameter, is unapproached by any other record of icebergs in scabland floods. Mounds are largest and most numerous close to their upper limit, 850 feet A.T. Above that are only stray erratic boulders, up to 1100 feet A.T. Two episodes of berg flotation thus seem indicated. Because even the massive bergs required for transporting the material of these larger piles would hardly ground in 225 feet of water, a prompt subsidence of the flood while bergs still survived in the Cold Creek ponding seems required. The abundance of bergs, the enormous size indicated, and the probable brevity of the episode appear to demand special conditions for release from parent glacial ice (a bursting Lake Missoula dam?) as well as special conditions for their transportation. The lithology of fragments in the till (Bretz, 1930b, p. 410) strongly suggests derivation from Beltian terranes of northern Idaho and western Montana. No route from Spokane across the scablands to Cold Creek valley could avoid cataracts and cascades where break-ups would be probable. It seems that the delivery of great numbers of giant bergs to Pasco basin must have required by-passing of the scabland river routes, therefore a trip along an unblockaded Columbia. Air and water currents probably caused their collection on the west side of both Pasco and Quincy basins.

The largest bar gravel deposit in Pasco basin lies a few miles upstream and across the Columbia from the Cold Creek bar (Fig. 17). It covers most of the northern half of the Priest Rapids quadrangle, and its contours nicely fit the inside of the uniform 90-degree curve made by the river. A summit area of about 20 square miles is enclosed by the 800-foot contour and reaches 860 feet A.T. more than 400 feet above the river. Except for the Columbia River undercutting along its southern margin (a 200-foot cliff) and at its upstream contact with Saddle Mountains (a 100-foot bluff), there are essentially no marks of erosion on its smooth, rounded form. No better place for survival of terraces cut during removal of a gravel fill in Pasco basin can be asked for. Yet two traverses and the 10-foot contours of the U.S.B.R. map reveal no such forms.

A pit on the 800-foot contour near the north end of this Priest Rapids bar shows a coarse, almost entirely basaltic gravel. Columbia River

FIGURE 17.—CENTRAL PART OF PASCO BASIN

Contour interval 100 feet, locally 50 feet. From Connell, Coyote Rapids, Eltopia, Hanford, Pasco, Prosser, Scooteney Lake, and Wallula topographic maps, U. S. Geol. Survey.

gravel immediately to the west is only 50 per cent basalt. Thus the glacial Columbia was carrying plateau, *i.e.* scabland, waste almost exclusively when the huge deposit was made. Sorting and wear of this gravel indicate sources more distant than walls of the Saddle Mountains water gap, 3–4 miles upstream. The gap therefore was already well opened and deepened when the massive gravel deposit was made.

A roadside cut near the southern end of the summit flat showed southeast-dipping fore-sets. A few borings show sand, gravel, and cobbles in the upper part of the bar but only gravel and boulders in the basal portion. At depths of about 100 feet these rest on Ringold clay and sand.

A.E.C. restrictions prevented examination of the southern cliff. If the deposit is a bar, stratification here should be either (1) steeply east-dipping, delta-type fore-sets, or (2) gently southwest-dipping beds parallel to the slope of the bar, with shallow fore-sets dipping into the bar, as in the Lind Coulee bar at the East Low Canal crossing. Construction of the Priest Rapids dam on the Columbia, now authorized, should resolve this question.

If this massive deposit be considered an erosional residual, the 18–20 square miles of lowland immediately beyond its east-facing scarp must be explained. That tract has no stream, and it cannot possibly be a meander scar made by the Columbia. Even the 25-foot contour intervals of the Geological Survey maps show clearly that it is another Cold Creek Valley in origin, an unfilled fosse behind a valley-margin bar.

This Priest Rapids bar is oriented and streamlined for the same great river that made the Cold Creek and Gable Mountain bars and in but one place shows signs of any smaller bar form superposed on its smooth riverward slope. Its thickness is approximately the height (75–125 feet) of the 4-mile-long lee scarp. The low area immediately east of it, underlain by Ringold, is obviously a remnant of the original surface on which the bar was deposited.

A smaller and lower bar, product of a later and smaller flood, trails about 4 miles downstream from the cliff in the huge bar's southern extremity. On the U.S.B.R. map, this younger bar top (125 feet above the river) is shown 60 feet higher than the dune-covered fosse between it and the low land east of the great bar. Yet the low area is 100 feet higher than the top of the younger bar.

Both the Cold Creek and Gable Mountain bars, separated by an undrained sag, can be identified farther south by linear elevations in the dune-covered central tract of Pasco basin (Fig. 17). The fosse of the Cold Creek bar determines the course of the creek to its junction with Yakima River.

Another conspicuous basin bar, also with a dune sand cover, lies just east of the Columbia and northwest of Pasco (Fig. 17). Its top is 200 feet above the river, its fosse 100 feet deep. Esquatzel Coulee's western channel enters the fosse, hanging 40–50 feet above its floor. Were the bar contemporaneous with the channel, the last 2 miles of the tapering form could hardly have been deposited.

These valley forms in the Pasco basin were never made erosionally during normal dissection of a complete valley fill.

The formidable problem of removing a fill in this basin, 16 miles wide between the 750-foot contours in the latitude of Pasco (Flint asked for the 900-foot contour), was handled by both Allison and Flint in the most general terms, with no comments on the surprising absence of extensive terrace remnants. Lower Yakima structural valley is another large basin which must have shared in this fill (Pl. 1). Its width is much greater in proportion to its stream than Pasco basin, but it likewise is empty and is unterraced except close to the river. It is quite unlike the strongly terraced Yakima Valley above Ellensburg. Flint and Allison said nothing about the place of lower Yakima Valley in their theoretical fills.[27] It is

[27] Flint ignored the scabland and upvalley fore-sets in associated gravel bars described and figured by Bretz (1930b, p. 412) in the Chandler Narrows at the mouth of this structural valley; and made no text comment on the pebbly silts veneering Yakima Valley slopes up to 1100 feet A.T. Allison (1933, p. 678), placed an ice jam in the Narrows which, by growing higher, detoured the upvalley flood current onto successively higher surfaces and thus made possible the 450-foot vertical range of this very local scabland. Unaccountably, however, the lower (earlier) bars are still perfect. The superposition of the growing dam did them no harm.

Walla Walla structural valley, southeast of the

argued here that no such fills ever existed, and it is suggested that the supposed correlatives below Wallula Narrows are probably older deposits shallowly covered with fresher debris. Huge gravel mounds comparable to the scabland bars have been described (Bretz, 1925; 1928a) along the Columbia as far as Portland. No restudy of that region was possible during the 1952 field season.

Allison's Snake River gravel bars, although older than Flint's terraces, are also relics of former valley fills. They cannot be remnants of a Snake valley train. Consistency in the ice-jam hypothesis therefore should postulate a comparable prescabland aggradation in major valleys throughout the plateau with local thicknesses equal to heights of its largest bars. As they are prescabland, these valley fills must antedate the divide crossings. This, with their overwhelming percentage of basaltic and caliche gravel, requires local origin on the plateau. Thus plateau valleys unentered by glacial water should still possess records of this episode. None does. If the plateau scabland bars are deposits of glacial streams, the Snake Canyon bars should be accepted as of the same genesis.

The high Wallula scabland, 800 feet above the river and more than a mile wider than the top of the gorge, was accepted by Allison and Flint as channeled and accompanied by scabland gravel deposits. Although extraordinarily high, its altitude is that of the Yakima Valley and Walla Walla basin pebbly silts and drifted erratics and approximately that of upper-limit records along the lower Snake. To explain these facts, Bretz (1925, p. 257) proposed that Wallula Narrows had acted as a bottleneck for the combined volume of all scabland rivers, and even announced the discharge through the short canyon as 39 cubic miles per day. This proved to be the most shocking of all interpretations based on the flood hypothesis, and Bretz retreated subsequently, taking refuge in possible postscabland anticlinal uplift or postscabland (and late scabland) deepening at the Narrows, for neither of which is there any known field evidence.

The scabland on top of the Wallula gorge walls apparently must be early, although it hardly suggests its age. It would seem to predate the development of the Quincy basin cataracts which, 100 miles distant, have brinks only 100 feet higher. It may be correlative with their earliest westward discharge at 1360 feet A.T. Allison's or Flint's fills would readily resolve the problem if only there were supporting evidence for them. But even the concept that scabland bars are erosional forms developed in former valley fills cannot be used with Pasco basin's and lower Yakima Valley's great empty widths confronting us. Nor can the high Wallula scabland, when compared with Drumheller, Othello, Palouse-Snake divide, Grand Coulee complex, and other contributing scabland, possibly be adequate for the ice-jam "run-around" concept. There may be some possible combination into an eclectic hypothesis which future study will bring out. If only one factor be involved, the bottleneck idea again seems most acceptable. At the rate computed in 1925, Lake Missoula's 500 cubic miles of water would pass through the Wallula Narrows in a little less than 2 weeks.

Palouse-Snake Divide and Environs

(Maps: Haas and Starbuck topographic and geologic maps; U.S.G.S.)

(Bretz and Smith)

General

Glacial water, flowing south down the Cheney-Palouse tract, was in excess of the amount Washtucna Coulee could carry, and it widely overflowed a structurally determined

Columbia-Snake junction (Pl. 1) also has (1) gritty silts up to 1050 feet A.T. or more, (2) scabland-marked narrows separating it from the larger valley and (3) fore-set gravel containing foreign pebbles with dip upvalley away from the Columbia (Bretz, 1929, p. 531). Both men have ignored these features. No theoretical ice jam can explain this because the fore-set gravels lie on the floor of a marked broadening, 4–5 miles east of the narrows. In contrast with Yakima Valley, the Walla Walla basin contains thick deposits of silt and fine sand derived from a glacial Columbia, with marked contrasts between, and disconformities separating, different members. The basin appears to have repeatedly become a shallow post-flood lake. Detailed study of these sediments may yield significant data regarding the glacial Columbia's flood and non-flood experiences.

divide, escaping southward from Palouse River valley to enter the canyon of the Snake, making a scabland record on more than 80 square miles of the divide summit and producing the deep canyon (top width 1250 feet, maximum depth 440 feet) by which Palouse River now enters the Snake (Pl. 1). Interpretations of the conditions which caused this overflow and canyoning vary widely. Four new topographic maps of the U. S. Geological Survey (Benge, Haas, LaCrosse, Starbuck) and a complete coverage by aerial photography supply some pertinent new data.

Flint believed that any first-magnitude flooding across the divide would find, or make, a ponded Snake Canyon and would build "a large deltaic deposit whose foreset beds would have an amplitude of hundreds of feet" in the deep Snake River canyon (1938, p. 513). Because "nothing of the kind is found in the Palouse Canyon scabland", Flint believed that an earlier proglacial flow had occurred, finding "a very low route" already in existence across the Palouse-Snake divide to permit spill-over from "a comparatively thin fill" in the Palouse Valley. This gashed the divide to make the Palouse Canyon transection (1) before the major episode of filling occurred, (2) before backwater (Lake Lewis) began rising in Snake River canyon to cause the deposits at the Palouse-Snake junction which Bretz had termed bars, and (3) before any scabland had been made on the divide. These "bars" Flint viewed as erosional remnants of a slowly aggraded dam, comparable in character to the Chippewa River deposit in the Mississippi River which forms Lake Pepin. This dam (800 feet thick) formed "Riparia Lake" upstream in Snake River canyon. By coincidence alone, Lake Lewis was slowly rising at the same time, thus avoiding long delta fore-sets in the deposits. Such fore-sets, however, do exist. Flint saw them (1938, p. 479, 481) but did not explain them.

Scabland Features on the Divide

With reference to Flint's supposed "very low [preglacial] route", it is instructive to note that an east-west profile along the high northern part of this divide, drawn from the Haas and Starbuck quadrangle maps (Fig. 18), finds the entire 9-mile width on the summit scabland to be 1280 feet or higher, more than 200 feet above the present aggraded Palouse Valley floor, except for an eighth of a mile immediately adjacent to the canyon walls on either side, where it is 1200 feet. A mile south of the cross section the brink of the canyon walls is 1280 feet. The loessial scarps marginal to this cross section are 120 to 180 feet high, and at least 100 feet of loess probably covered the divide top when the first glacial spill-over occurred. There is also at least 50 feet of gravel in the bottom of the preglacial Palouse at this place. These facts require a thickness of not less than 350 feet (to 1325 feet A.T.) in Flint's "comparatively thin fill" to make the first crossing possible. His "very low route" never existed, and Palouse Canyon cutting belongs in the general scabland-making episode.

Flint minimized the size of the various cataract alcoves of this district, citing five examples on the divide, not one of which is deep enough to be shown on the new topographic maps. His argument collapses when the map of H U Ranch dry falls (Fig. 19) (Haas quadrangle S. W. ¼ Sec. 29, T 14 N., R 36 E.) is examined. The cataract cliff is 280 vertical feet from the lowest notch in the lip to the bottom of the plunge pool. Recession of the falls left the Davin gorge 1½ miles long containing a closed depression equally long and 128 feet deep.[28] The water that supplied the H U Ranch cataract even at its very last functioning had to cross the but slightly channeled scabland divide at an altitude of 1280 feet A.T. or more.

Inspection of the Haas and Starbuck quadrangle maps shows several such features far exceeding in magnitude the examples cited by Flint. One of these (Devils Lake, crossing Sec. 17, T. 14 N., R. 37 E.) has an 80-foot dry falls at the head of a recessional gorge about 2 miles long, more than 100 feet deep, and containing closed basins with 40–80 feet of closure. The largest and deepest of these terminates almost on the brink of Palouse Canyon, itself 200 feet deeper at this place.

[28] One of Flint's examples (1938, p. 490), is just west of the Davin Ranch where one may look northward up the gorge to the cataract cliff.

Another dry falls (S.E. ¼ Sec. 1, T. 14 N., R. 36 E.) tunneled through by the Union Pacific R.R. between canyon and preglacial valley is 100 feet high.[29]

Were it not for late development of a short channel from the northeast, 80–120 feet deep above the H U Ranch cataract and thus an incision of its lip, the sheer drop here would

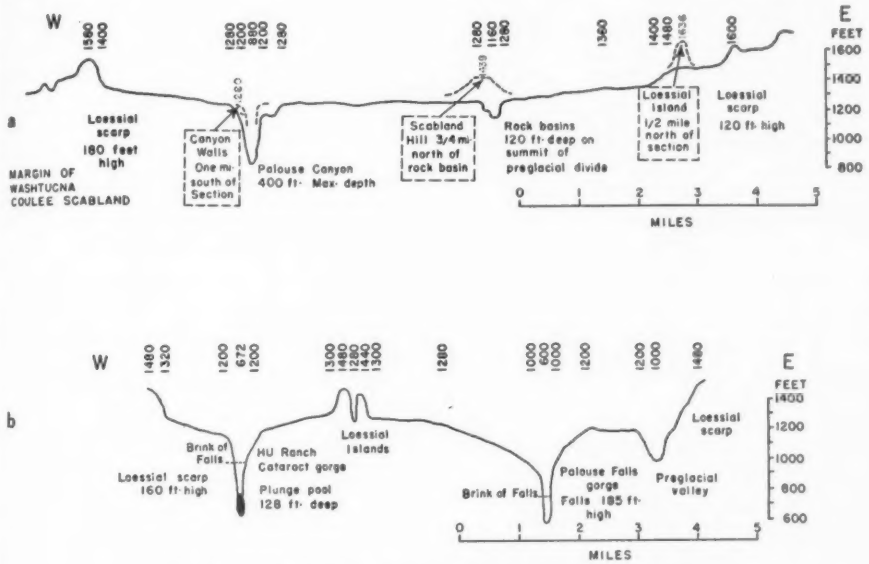


FIGURE 18.—PROFILES OF SCABLAND ON PALOUSE-SNAKE DIVIDE
a—Along the crest of the former divide
b—Immediately downstream from the two cataracts

Flint also minimized the depth of rock basins in this district, as elsewhere in the area of his study. He could never have seen the two basins 1½ miles directly south of Hooper Junction on the very summit of the divide scabland (Sec. 3, T. 14 N., R. 37 E.) (Fig. 20). The larger one is ringed by a 1240-foot hachured contour and is more than 120 feet deep, and the smaller one is more than 80 feet deep below the same contour. These great empty holes are impossible erosional products of Flint's streams that removed the fill. If they were pre-fill in age, they would now be full of gravel. They cannot be forced into the fill-and-cut strait jacket.

[29] Trimble (1950) noted a striking joint control of the shapes and orientations of many rock basins, "slots" and gorges in the Palouse-Snake divide scabland, some closed basins 100 feet in depth. He accepted them as scabland features, the erosional results of glacial water, but disclaimed any "attempt to explain the mechanics of their erosion."

be 350–400 feet. Figure 19 suggests that before this channeling occurred the cataract head was a cascade chute at the northwest end of the gorge. Alignment of this chute with the straight recessional gorge is determined by a vertical fault with 40 to 50-foot downthrow to the east. But the chute head is close to the western margin of the divide scabland, and the larger volume of water came from the northeast. The lengthening channel, therefore, left the fault trace and headed 2–3 miles to the northeast. The profile of the west wall of the Davin gorge shows that the cataract became markedly higher as it receded along the last half mile of its length by migrating out of a prescabland, minor valley along the fault.

Bases of the loessial scarps bounding the Palouse-Snake divide scabland are 1400 feet A.T. or higher on both sides at the north end of the wide spillover. Flint got the vertical range of 120 feet or more in summit scabland

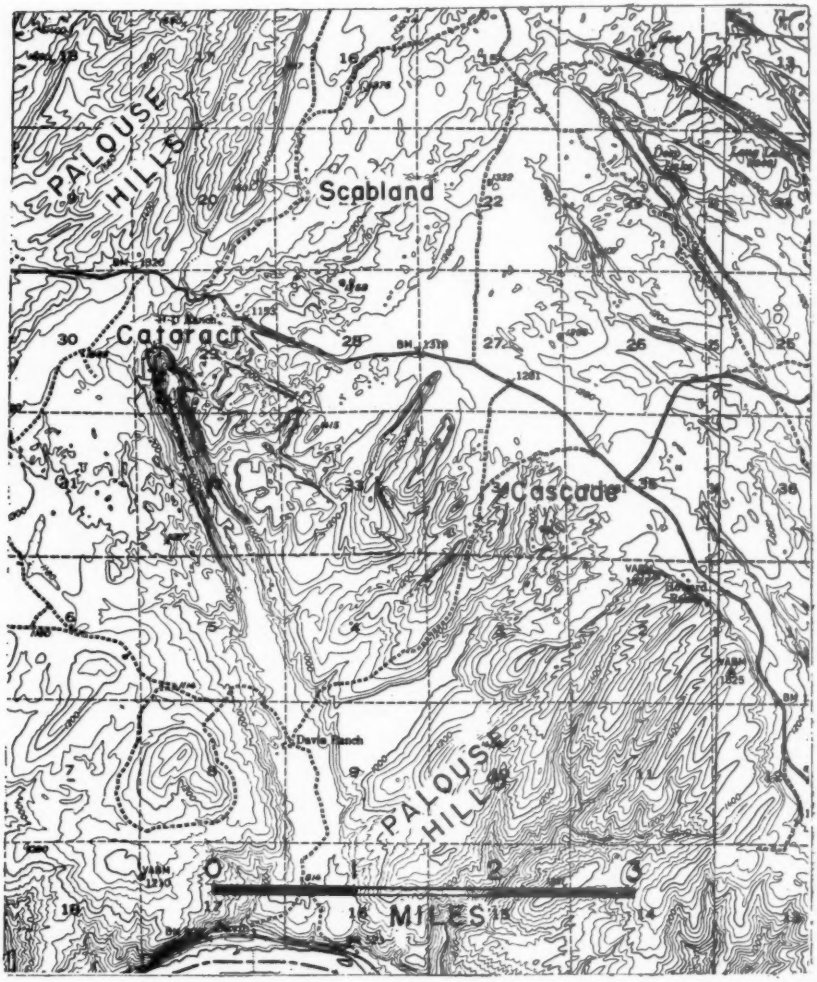


FIGURE 19.—VICINITY OF H U RANCH CATARACT

Western part of Palouse-Snake divide scabland. Loessial topography in northwestern and southeastern portions of area, loessial islands in central part. Largest island is 1 mile long and 500–1000 feet wide, and its scarps are 150 feet high. Deep Lake gorge is tributary to Palouse Canyon a mile beyond eastern edge of map. Part of Haas and Starbuck topographic maps, U. S. Geol. Survey. Contour interval 40 feet.

by aggrading an equivalent amount so that a river whose "discharge was less than that of the Snake today" could reach successively higher and wider surfaces. However, his river did not make the scabland, it only scoured off the loess. The scabland relief was produced later, while the fill was being removed, and this occurred simply because original preglacial gradients were being restored, not because of any augmentation in discharge. If a normal valley train already existed in Washtucna, it could have been a factor in causing overflow, but the 80 square miles of scabland on the preglacial Palouse-Snake divide can never be explained by Flint's theory. The acid test, however, is in features on the south side of Snake River canyon.

Allison never specifically located any ice

jam on the plateau scabland but was content with general statements regarding what would happen if they did form in a drainage system invaded by meltwater streams. For the Palouse-Snake divide's scabland, he would obviously have to block Washtucna Coulee (Fig. 2; Pl. 1) somewhere along its length to a height great enough to cause the first spillover, estimated as at least 300 feet above the present aggraded Palouse Valley floor, or 1325 feet. A.T. However, water reached an elevation of 1400 feet at the bases of the loessial scarps margining this 9- to 10-mile wide tract, and Allison's theory requires more ice jams not alone in the growing Palouse Canyon but also on the divide top itself to avoid the great volume. Where can they be placed on this great width so that water would have to rise to detour around them, make the higher scabland, and cut the loessial scarps? To the writers, such a mechanism is impossible here. Only in constricted or crooked valleys could jams ever grow to the required dimensions. Furthermore, in the Washtucna Coulee not a single favorable place was found for the jam called for nor any features like those Waters (1933, p. 818) interpreted as ice-jam records in Columbia valley north of the plateau. Still further, the loessial scarp bases along the 35 miles of Washtucna's length constitute a uniformly descending profile, impossible to explain if an ice jam interrupted the glacial stream's surface gradient.

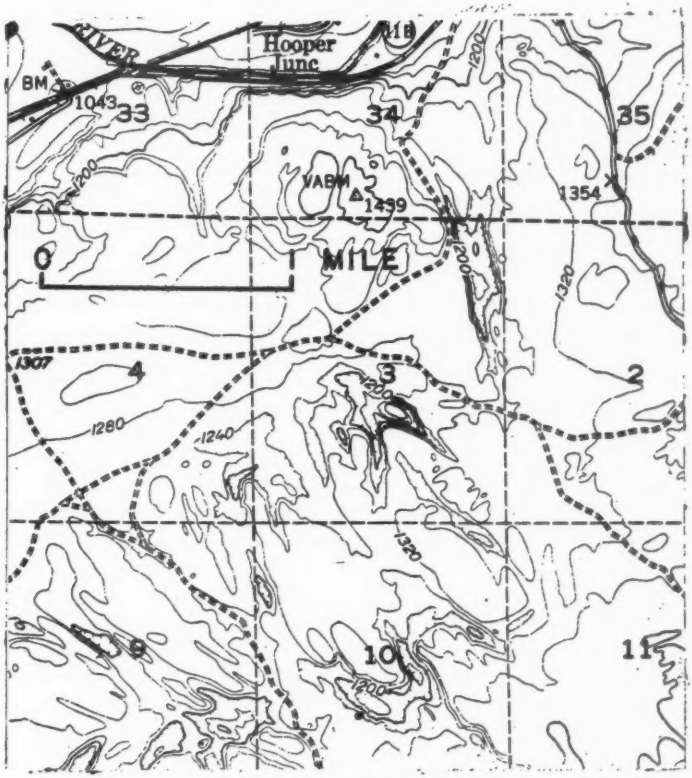


FIGURE 20.—ROCK BASINS ON SUMMIT OF PALOUSE-SNAKE DIVIDE

Basin in south-central Sec. 3 is more than 120 feet deep. Vertical range of scabland in this tract is 400 feet. Part of Haas and Starbuck topographic maps, U. S. Geol. Survey. Contour interval 40 feet.

To erode Palouse Canyon across the divide deeply enough to provide a permanent rerouting of the river, any ice jam in Washtucna

must have persisted (or been repeatedly renewed) while Allison's moderately sized, detoured meltwater river deepened more than 400 feet in basalt. The narrow new route was never (more than temporarily) clogged with the abundant floating ice while the dam in the much wider preglacial valley continued to be effective. How can rock basins 100 feet deep on the scabland summit of the divide be explained by Allison's hypothesis? How could buttresses of loess for the divide top's dam survive? How, when buttresses failed, could the dam grow laterally to close the gaps seized upon by the escaping water and thus eventually make the 9–10 mile width of summit scabland? The quantity of floating ice constantly required of a normally wasting ice front to maintain and extend a dam on this broad summit passes all credibility. Failure of a Lake Missoula dam might rapidly release the enormous quantities of berg ice demanded by this elaboration of Allison's concept, but no believer in moderately proportioned glacial rivers will ever accept that solution of the problem.

Bars in Snake Canyon

(Bretz and Smith)

At the Palouse-Snake confluence, Allison and Flint have reinterpreted the "bars" (Bretz, 1928a), disagreeing with each other as well as with Bretz. Field study in 1952, with the aid of the Haas and Starbuck topographic maps, has led to a vigorous reaffirmation of the original interpretation and a denial that any dam of debris or ice ever existed at this place in Snake River Canyon.

Bretz originally described two great bars at the junction, both on the south side of Snake River. One was a "mid-canyon bar" about 2 miles long on the floor of the canyon, deposited by glacial water flowing tumultuously *up* the Snake. The other was a high-lying "shoulder bar" in the mouth of Fields Gulch, a tributary from the south entering the Snake about 2 miles farther downstream. The shoulder bar was made by a detour of part of the flood southward across the Snake and thence westward around the south side of a big, "semi-isolated" basalt hill, there to reenter the Snake in the south wall of its canyon (Fig. 21).

Flint interpreted these bars as remnants of his Riparia Lake dam, a Snake canyon fill 800 feet thick, actually the toe of his 120-foot fill on the top of the Palouse-Snake divide. Lake Lewis, the cause of all this valley deposition, was supposedly rising downstream in Snake River canyon at the same time, and the rate of deposition at the dam site kept pace, so this dam never developed delta structure.

Flint was a bit reserved about the composition of the mid-canyon bar, saying of the "rubbly, basaltic, cobble gravel with huge boulders, too little size-sorted to have distinct bedding" that "it strongly suggests torrential conditions" (1938, p. 513). The depression (fosse) between the bar top (241, not 150, feet above the Snake) and the basalt knobs and cliffs south of it, he said was "unusual" (p. 481) but appeared to be comparable to " 'deeps' at the outside of bends in the river during dissection" of his dam.

Allison (1941, p. 69–70) said of the mid-canyon bar that its "top is constructional", its "flutings are primary", and the depression back of it is also primary, showing "no resemblance to deeps at the outside of river bends nor to plunge pools". "It represents the top of the fill made by the last outpouring of glacial waters from Palouse River Canyon". Although that "outpouring" of "torrential" water had to be *up* Snake River canyon, neither Allison nor Flint noted the fore-set beds at the eastern end of this bar. They dip *up* the Snake and toward the depression.[30] Allison thought the depression "was left unfilled—because the area was occupied by ice, driven against the cliff by force of water from Palouse River canyon and then protected from wasting by shadow". There is, however,

[30] With Lupher (1944), they have ignored the existence of such up-valley fore-sets, reported by Bretz to occur as far as Lewiston. Existence of the new Hay geological map (40-foot contour interval) and of State Highway 3 should encourage examination of the readily accessible Central Ferry "half mound bar" (Bretz, 1929, p. 411), 20 miles up the Snake from the mid-canyon bar. Despite the authority of the map, it cannot be correctly described as a Snake River terrace.

FIGURE 21.—SCABLAND FEATURES ON SOUTH SIDE OF SNAKE RIVER AT CONFLUENCE OF PALOUSE RIVER
Fields Gulch preglacial course shown by dashed lines. Part of Starbuck and Haas topographic maps, U. S. Geol. Survey. Contour interval 40 feet.

no summer shadow on the area from adjacent cliffs.

From the mid-canyon bar minor dependent bars extend southward, *away* from the Snake and back in the mouth of a blocked, preglacial tributary that formerly entered the canyon across the north edge of section 6, T. 12 N., R. 37 E., and now is detoured westward to Fields Gulch. The headlands at the former mouth are marked scabland buttes rising above these southward extensions of the mid-canyon bar. Each butte has a long gravel ridge in its lee that tapers out and down *away* from the Snake. These gravel ridges are genetically part of the mid-canyon bar. Their existence, magnitudes, and space relations must not be ignored or slighted. The knobs reach 800 feet A.T., and the south-pointing bars are as high as the mid-canyon gravel hill, 731 feet A.T.

From Lyons Ferry at the junction of Palouse and Snake rivers (Starbuck quadrangle) a circuitous road may be followed to DeRuwe Ranch (Haas quadrangle), just south of the shoulder bar in Fields Gulch (Fig. 21). In the second and third miles of this road, one climbs from BM 547, entirely on nonterraced gravel, to BM 861 in the detoured drainage. Enroute, constructional forms are very obvious in the moundings and enclosed, undrained depressions. The striking little "trickle bar"[31] (Pl. 15, fig. 4) which descends the southwest slope of a scabland hill in SW ¼ Sec. 36, T. 13 N., R. 36 R. looks somewhat like a flattish debris cone or a steep detrital fan, but it is definitely ridged along its length, its lateral slopes are much too steep to harmonize with the gentler gradient along the crest, and its bulk is far in excess of what could be shed from the small section of the basalt hill that could contribute. The ridged fan or cone form mounts almost to the hilltop. The altitude of its crest ranges between 920 and 1000 feet. The deposit originally blocked local drainage off the hill. A gully across its low distal end has spared portions of the floor of the once-closed depression back of the bar. The valleyward lateral slope at the terminus descends 50 feet and flattens out in a 40-acre closed depression shown on the Starbuck quadrangle map. The deposit is composed almost entirely of little-worn basalt pebbles and boulders. A few well-rounded Snake River pebbles were found. The form is that of a small, local bar 390–520 feet up on the bar-mounded slope above Snake River but 300–430 feet below the level of Flint's supposed dam across Snake River canyon. The water that carried its debris southward down this hillside away from the Snake came over the top, 1040 feet A.T.

The shoulder bar and immediate environs (Fig. 21) were carefully examined during this field study. Renewed excavation in the old pit on the north side afforded good sections, and the topography of the entire deposit was seen by walking over it. The interpretation that it is a great gravel deposit essentially with original outlines and with long deltaic fore-sets dipping northward toward the Snake is here emphatically reaffirmed.[32]

The flattish summit of this shoulder-bar deposit is approximately 1000 feet A.T. over nearly half of Sec. 25 and more than half of Sec. 26, T. 13 N., R. 36 E. It is diversified by low linear ridges and swales, their relief 10–15 feet in places, impossible to explain as erosionally made by subsequent runoff. A few well-worn Snake River pebbles were found; otherwise the material is all little-worn basalt.[33]

Recognition of the true character of this minor relief on the summit (Pl. 13, Fig. 1) did not come until aerial photographs were studied after the writers had left the field, hence no measurements were made. Scaled from the

[31] Long narrow bars steeply descending preexisting slopes and appearing as trickles from a distance.

[32] Among Allison's comments on Flint's work is the statement that this shoulder bar "was built into water [525 feet deep] at the edge of Snake River Canyon by currents which left the major valley a short distance, [2 miles] upstream [and detoured around the south side of the semi-isolated hill, 1241 feet A.T.] when free flow down the valley was obstructed by ice jams."

[33] Allison's view that the gravel of the shoulder bar is largely reworked, old, high-level Snake River terrace gravel is untenable. The percentage of well-worn constituents is small, of nonbasalt fragments even smaller. Furthermore, Snake River canyon for nearly 75 miles upstream, where Allison's supposed old, high-level terrace system should be well recorded, has no such deposits at comparable altitudes above the river. The 20 miles of this canyon shown on the Hay and Starbuck geological maps (U.S.G.S. 1954) have but one patch of gravel (Qs-Quat. scabland) as high as 450 feet above the river.

photograph (Production and Marketing Adm., CCH-16-197), the crests of the giant current ripples are about 175 feet apart.

From this summit flat, the bar surface descends 200 feet southward and westward with a convex slope which the Haas quadrangle map shows as crenulated. Orientation of the ridged crenulations changes gradually from nearly south-pointing forms on the south side to west-pointing, even northwest-pointing, forms on the west side of the hill. A little gullying has occurred between the ridges low on the slopes, but, for the most part, they are original shapes of the bar surface.[34]

Sections in the old railroad gravel pit on the Snake River face, cut deeply back into the deposit, show the bar material to be fore-set bedded almost throughout (Pl. 14, figs. 1, 2, 3). Parallel to the steep, riverward slope of the gravel hill, these long fore-sets are not a misinterpretation of pseudo-fore-setting due to slopewash. The 100-foot scarp trending northeastward from the pit for a quarter of a mile is the constructional front of the deposit, avalanching as in a dune. It is not an erosional scarp. Its strike parallels that of the long fore-sets in the pit (Flint, 1938, p. 481, to the contrary notwithstanding) and the elongation of the giant ripples. Water, not debris, or ice, filled Snake Canyon up to more than 1000 feet A.T. at this place when these fore-sets were deposited. Flint offered no explanation for them.

In this fore-set gravel is embedded a granite boulder 5½ feet long, its outlines fracture-determined and only one surface showing pre-fracture abrasion. Flint (1938, p. 498) remarked that "In such shallow scabland streams as are indicated by the stratification of their deposits, only small pieces of glacier ice could be expected to travel down their lengths. . ." How small a piece of glacier (or river) ice could have held onto this granite boulder for the 75 miles it travelled down the Cheney-Palouse scabland river? And meandering added to that mileage![35]

This shoulder bar in the lower part of Fields Gulch *contains the delta fore-sets Flint asked for* at the mouth of the Palouse, if the flood theory were correct! The mid-canyon bar is part of that "delta"; the bars tailing southward from the scabland buttes in sections 30 and 31, T. 13 N., R. 37 E., are part; the bars along the road south from Lyons Ferry are part; the "trickle bar" is part; the bars farther west, tapering southward from the same hill are part. The flattish summit of the shoulder bar with its giant current ripples is the *subfluvial* "delta top". All gravel forms are portions of a deposit made at the *bottom* of the great torrent which carried most of its debris *across* Snake Canyon and much of it *up* for more than 500 feet on the far side. Volume and velocity of the flood that made these bars were greater than anything that Flint objected to.

Allison, accepting this remarkable detour of glacial Snake River, made it possible by placing an ice jam in the canyon just below the mouth of the Palouse (Fig. 21). A slight constriction here might favor the idea, but the dam would have to be from three quarters of a mile to a mile wide (across the canyon), and, to allow the detoured stream to re-enter the canyon where it did, the dam could be only 2 miles long. Such a dam never could have had enough frictional grip on the canyon walls or among its constituent ice blocks to hold back a 500-foot head of water. Furthermore, there was equally deep water where and when the detour re-entered the canyon below his dam. Above the ice jam, there was only slack water, and through this the scabland gravel had to travel to cross the canyon and climb to the shoulder-bar summit. Allison's hypothesis really requires a scabland run-around canyon at this place, instead of a gravel deposit.

Allison needs another ice jam for his sequence of events at the Palouse-Snake junction. Plate 1 and Figure 19 show that the 4-mile wide Davin Ranch portion of the divide scabland, lying west of the isolated group of loessial hills, had to be effectively blockaded to make the detour south of the Snake possible. Otherwise, all

[34] A repeating character of the giant current ripples seen during the 1952 field season is the extension of the ridged forms of a bar summit down the slope toward the adjacent channel. Neither slopes down into a fosse nor erosionally scarped valleyward slopes have shown this feature.

[35] Another granite boulder, equally large, in the Cow Creek scabland 1½ miles north of Hooper lies as high as the scabland on the divide. Still another and larger (9 feet maximum diameter) granite boulder lies 3 miles south of Ewan, 35 miles down along the Cheney-Palouse tract.

Cheney-Palouse glacial water would have entered the Snake below the site of Allison's canyon dam, and only Snake River water would have used the detour. Coming directly from a ponded Snake, it could have had no gravel load, and no shoulder bar could have been made.

Prescabland Fields Gulch

(Bretz and Smith)

The floor of Fields Gulch in the half mile between the shoulder bar and Snake River is very narrow compared to its proportions farther upstream (Fig. 21). It is a hanging-valley mouth, with a sharp entrenchment made subsequently. Its gradient is 200 feet to the mile in this stretch, six times as steep as in the next 5 miles upstream. It clearly does not record the stream that preglacially drained the 50 contributing square miles of mature topography lying to the south. Even if a partial damming of the gulch by the bar (Bretz, 1928a, p. 659, Fig. 8) interrupted the gradient, this narrowness and steepness would indicate that these great bar deposits have blocked a capacious lower course elsewhere, and have turned postscabland drainage into the present aberrantly narrow and steeply graded route. A reconstructed prescabland Fields Gulch (Fig. 21) would enter Snake Canyon where the mid-canyon bar and other higher bars obviously fill and block an earlier drainage way to the master stream. Compared with the valleys of neighboring Tucannon River and Kellog Creek, the width here between basalt hills is certainly adequate for the drainage area. And this blocked valley is twice as wide as the present one downstream from the shoulder bar.

Flint's Riparia Lake

(Bretz and Smith)

For the following reasons, the writers hold that Flint's Riparia Lake never existed.

(1) All roads crossing the region where the distal portion of Flint's 1300-foot dam (Fig. 21) must have lain were traversed, and a few miles of foot traverse was made across interroad tracts. No remnants of any gravel or sand deposit were found on the broad, thinly loess-covered, basalt uplands above the summit level of the shoulder bar, and no suggestion of debris from such deposits was seen in drainage ways. It is quite impossible that all that theoretical deposit could have been carried off these hilltops where it was originally 100–300 feet thick.

Only wind, rain, and slope wash are available agents. Had they been thus effective, there would be no remnants of Flint's fill elsewhere, nor of Bretz's bars, to theorize about.

(2) The Haas and Starbuck geological maps (U. S. Geol. Survey, 1954), made for engineering use in proposed dam construction on Snake River, show no gravel deposits above about 1100 feet A.T. on the area covered by Flint's hypothetical 1300-foot dam. The highest part of the shoulder bar's gravel is 1120 feet. Along the east side of Flint's fill on the Snake-Palouse divide, altitude of which he showed as 1350 feet, the Starbuck map shows a ragged pattern of local drainage gulches in basalt, with loess-covered interfluves (but no gravel remnants) descending to and below 1200 feet A.T.

(3) The giant ripple marks covering the shoulder bar's square-mile summit flat record a constructional surface, the original top of the deposit, 325 feet below the altitude of Flint's delta surface at that place.

(4) Flint's Figure 9 (1948), showing the delta-dam gradient as about 13 feet per mile, also shows a remarkable steepening downstream to 30 feet per mile. The distal margin of the supposed dam had to be at least 1300 feet A.T. for 10 miles west of the Palouse junction in order to impound Lake Riparia. The upper limit of Snake Canyon flooding 5 miles farther downstream is only 1200. (*See* Flint's Fig. 11.) Hence the incredibly steep gradient!

(5) Half a dozen gravel-covered tracts of 160 acres or more, lying on the Palouse-Snake divide scabland, are shown on the Haas and Starbuck geological maps. Two closely associated ones, just east of Palouse River about 2 miles above its mouth (Starbuck map), partially bury interfluve spurs of the preglacial basalt topography. Where nearest each other, they are separated only by a small streamway draining from higher loess-covered country farther east. Two tributaries join, just above

this narrow place. Their 75-degree divergence upstream (eastward) outlines another basalt interfluve spur of comparable size, lying between the two. As shown in Flint's Figure 3 (which has some considerable altitude errors) this spur has no gravel cover. The map distance between the two gravel patches is a tenth of a mile at the west where the tributaries join and increases upstream to three quarters of a mile across the gravel-free spur.

"A single course of fore-sets [at least] 35 ft. thick" (Flint) in the northern deposit, described by Bretz (1928a, p. 655) and confirmed by Flint (1938, p. 479), dips N. 68° E. Bretz read this upvalley dip, away from the Palouse, as recording introduction of the pebble gravel diagonally across the top of the buried spur and considered both deposits as essentially bars, their form and location conditioned by the underlying topography and direction of current. To Flint, they were erosional remnants of his Snake Canyon dam whose flattish tops (terraces) now lie 125–285 feet below its original surface at this place.

The outstanding objection to Flint's interpretation here is the impossible survival of the closely contiguous "terraces" which, down in the united stream valley, 290–330 feet below the lowest portions of the flattish tops and 600 feet below the original surface of his fill, are only 500 feet apart. By his interpretation, half a square mile of original gravel cover for the scabland spur, 125 to nearly 600 feet thick, has been removed through that constricted gap without widening it.

(6) Flint's Riparia Lake received the Snake River of that time as well as glacial water from the Palouse. Its outlet must have been a considerable river. Nothing in Flint's text deals specifically with this outlet. His Figure 9 has an arrow marked "Lake outflow" near the southern margin of the delta. Traced on the Haas and Starbuck geological maps, it traverses a basalt topography ranging from 840 to 1200 feet altitude! The arrow therefore indicates an entirely hypothetical channel across the vanished delta. Figure 9 also shows four "conspicuous abandoned channels" across the site of the dam, two of them about a mile back from its low southern edge, the others even farther back. One does not exist. Two are misinterpretations of unaltered, preglacial saddles, related to his lake only in that they lie below its level. The fourth (in SE ¼ sec. 28, T. 13 N., R. 36 E.) is truly a small scabland discharge route through a preglacial saddle at 1000 feet altitude, a mile northwest of DeRuwe Ranch and 3 miles back from his delta's distal margin. It is 325 feet below the top of the supposed dam at this place. There are good minor bar forms at intervals from the aggraded flat at the ranch buildings (800 feet A.T.), (which also has bar forms) up to the col. There are even unfilled depressions close to the summit, compounded of small scabland knobs and bar knolls. Obviously, a current of flood water crossed this saddle at 1000 feet, a part of the overflow when the shoulder bar was being built. But obviously also it did not spill thence down the steep ravine toward the Snake. The scabland notch is not a channel and does not record any Snake River course during dissection of the dam. Furthermore, its small capacity falls far short of Allison's needs, should he try damming Snake Canyon below Fields Gulch but above the mouth of the preglacial tributary on the far side of the saddle.

(7) With a gradient of 13 feet to the mile for the Cheney-Palouse fill and the top-set beds of hypothetical Riparia Lake dam (Flint's estimate), the lake's outlet must have been forced to skirt the far southern edge of his Snake Canyon deposit. This Flint recognized. Snake River today, therefore, should be in a post-dam canyon superimposed on the basalt upland at that same place, and the preglacial canyon's dam should still exist. Palouse Canyon (made before his episode of filling) should still be buried in gravel, and Palouse River should be following its Washtucna Coulee preglacial route. Instead, the Snake has inexplicably flowed *up* the eastern front of his delta, the lake outlet climbing 30–35 feet *against* the delta's lateral slope to relocate and re-excavate the preglacial canyon 3 miles north (*see* Flint's Fig. 9) of the only place the glacial Snake could have crossed the delta when at its maximum growth. The irregularly hilly basalt topography shown in Figure 21 forbids any notion that the post-dam Snake could, while re-excavating, have slipped laterally back into its preglacial course. The hill summit between the DeRuwe notch and Snake Canyon is 1160 feet A.T. Once caught in the notch, the river never could have escaped.

(8) There are no shore lines for Riparia Lake,

even of the faintest detectable character, on the soil-mantled gentler slopes of Snake River canyon upstream from the dam, although there was ample opportunity for them to develop during the gradual trenching of that deposit at the mouth of the Palouse. Most of Flint's Cheney-Palouse scabland gravel was "leisurely" carried off during that same time.

(9) There are no lacustrine deposits on these slopes. There is, however, a widely distributed mantle of unsorted, unstratified silt containing abundant grains of fresh basalt with nonbasaltic pebbles and boulders extending up to 1300 feet altitude (Bretz, 1929, p. 407–418). No Riparia Lake type of ponding in Snake Canyon can explain the up-valley transportation for nearly 75 miles of this pebbly silt from its source, the Cheney-Palouse scabland river. Vigorous surging up the canyon is required.

(10) Snake River valley alongside and below the summit of the mid-canyon bar is only 1000 feet wide, and one bluff is all gravel. By the fill-and-cut hypothesis, this is all the widening the river has been able to do in the canyon bottom since Riparia Lake was drained. In contrast, Washtucna has a bottom width of 2000–2500 feet for many miles, made by a small divergent strand of glacial water. No postglacial stream has used the coulee. Harmonizing these facts with Flint's concept is quite impossible.

Devils Canyon

Map: Connell Quadrangle, U.S.G.S.

(Bretz)

Streamless Devils Canyon, 15 miles west of Palouse Canyon, transects the same Washtucna-Snake divide and is subequal in width and depth. Cataract or cascade recession here, as at H U Ranch falls, made the 4-mile gorge along a preglacial tributary of the Snake. The glacial stream made no scabland on top of the walls, and it ceased to flow before recession had completely cut through the divide. A rock sill at the canyon head, nearly 100 feet above Washtucna's floor, is tunneled through by the Spokane, Portland, and Seattle R.R. between canyon and coulee. Summit of the sill is more than 950 feet A.T., and the bottom of the canyon a mile to the south has an 800-foot depression contour. The canyon is empty thence to the Snake, but its flaring mouth has an unterraced, knolled, and ridged gravel mound rising nearly 300 feet above the Snake to the south and about 50 feet above a once undrained depression to the north. This mound projects more than a third of the way across the bottom of Snake Canyon and is entirely of gravel. The highest part stands nearly in midwidth of Devils Canyon. Figure 3 of Plate 9 shows a train of giant current-ripple marks half a mile long across it.

Flint derived Devils Canyon in part by falls retreat during removal of fill, originally 1300 feet A.T. at the canyon head. On this fill, the Devils Canyon distributary from the Washtucna distributary initially descended 100 feet in 4 miles (Flint, Fig. 11). Eventually it cut through about 400 feet of basalt at the head to leave the 150- to 200-foot cataract-cascade descending from the 100-foot sill. Then the canyon route was "suddenly" abandoned (Flint, 1938, p. 512). Thereafter, all Washtucna distributary water continued down along preglacial Esquatzel, subparallel to the Snake.

Downcutting in both the Snake and Washtucna-Esquatzel fills was controlled by Lake Lewis. At highest stage, the lake was only 21 and 25 miles distant. When lowered to Pasco basin altitudes, the subparallel routes were respectively 38 and 41 miles long.

To provide an adequate gradient for erosion of Devils Canyon, by falls or by uniform deepening, the Snake obviously had greatly to outrun Washtucna-Esquatzel in removing Flint's fill while, about the same distance down along each route, Lake Lewis was the common base level, and both streams had only gravel to erode. This is hardly an acceptable interpretation. That his Devils Canyon divergence ever could have continued deepening, once it encountered basalt, is likewise unacceptable. That the gravel mound semiblocking the canyon mouth is a terraced residue of either Snake or Devils gravel is denied by all its characteristics. It is another scabland bar. The empty and rock-bottomed gorge was indeed "suddenly" abandoned but by a stream far exceeding a distributary strand from a distributary strand of a river no larger than the Snake.

To use Allison's ice-jam theory specifically for Devils Canyon, we must (1) block discharge

across the Palouse-Snake divide, (2) construct another ice jam in Washtucna west of the head of Devils Canyon, (3) keep the narrow canyon head open while nearly 400 feet of deepening in basalt occurs, (4) have an ice jam down the Snake or Columbia higher than 500 feet A.T., and (5) remodel an older terrace into the gravel mound in the canyon mouth. These are difficult requirements.

Flint's Derivation and Disposition of Scabland Debris

(Bretz)

According to Flint (1938, p. 483), "very little" of the gravel fraction of the Cheney-Palouse glacial-river load was carried far enough down the Snake to reach Pasco basin. Deposition began promptly in Washtucna Coulee and Snake Canyon; the detritus being derived from "scoured basalt on steep gradients, mainly along preglacial drainage lines" (1938, p. 468) already down in basalt. These would have to be limited to Cow Creek, Rock Creek, Palouse River, and a few tributaries, and would involve but a small area compared with the nearly 1000 square miles eventually swept by glacial water in the tract of Flint's study. He visualized this fill as gradually extending upstream and finally reaching the channel heads. Because all channel-bottom basalt had by that time become buried, the channel-head deposits (required for his westward diversion of the northernmost distributaries to Crab Creek drainage) had to be supplied directly from the ice. Yet their "remnants" are 65 per cent basalt (Flint).

In Flint's succeeding episode of erosion of the fill, the rejuvenated glacial streams "notched and channeled the newly exposed basalt extensively" (1938, p. 468), making most of the scabland he studied. This erosional episode must have produced far more basaltic debris than did the earlier scouring. Most of the gravel already deposited, mingled with this newly made detritus, was carried across the Palouse-Snake divide to join the wastage from the 800-foot-thick Riparia Lake dam. Some entered Washtucna. All this gravel eventually went down along the Snake and Washtucna toward Pasco basin.

About 30 miles below Flint's dam site, the scabland upper limits (Secs. 20 and 21, T. 10 N., R. 33 E.) are 500 feet above the valley bottom. (Here Flint believed that scabland gravel graded into Touchet silts.) An average thickness of 650 feet for the 30 miles is therefore conceivable. A generous estimate of the average valley width involved is half a mile, thus a gravel-storage capacity of 2 cubic miles. If none entered Pasco basin, this represents all the basalt waste removed from nearly 1000 square miles of scabland. Evenly distributed over the Cheney-Palouse tract, it would account for about 10½ feet of erosion. Allowing for 10 per cent or less (Flint's estimate) of glacial drift in the gravel, and using Flint's porosity factor of 25 per cent (1938, p. 483), the amount of basalt eroded is about 7½ feet, barely enough to remove the weathered zone under the loess. By using Washtucna for additional storage and by leaving small "remnants" in the channelways, we may retain the estimate of 10½ feet. But this is much too small to produce the relief made by the scabland streams. One cubic mile of gravel distributed over the 80 square miles of scabland on the Palouse-Snake divide alone would provide a fill only 66 feet in average thickness[36] about half as deep as Flint's theory requires.

This computation assumes that Snake and Washtucna valleys were originally empty and that they received capacity fills. Nothing remotely approaching this quantity of gravel is in them today. Flint's theory requires removal down along the Snake as Lake Lewis subsided. Where has it gone if it did not enter and pass through Pasco basin? And how could it be carried across the Touchet deposits, during subsidence of Lake Lewis, without leaving a record? The complicated channeling in lower Esquatzel is not that recorded for Washtucna. Allison, finding Touchet beds only above scabland gravel, denied the contemporaneity

[36] The preglacial gradient of the Cheney-Palouse tract was 20–25 feet per mile. Flint's estimate for the slope of his fill was 13 feet per mile. Thus the deposit at Hooper (60 miles from channel head and 1112 feet altitude) had to be 420 feet thick (1530 feet A.T.) before any aggradation began at the head. By the time aggradation was completed, it was thus 275 feet thick on the divide summit (1280 feet A.T., 2 miles south of Hooper) and therefore had to lap an impossible 130 feet up on the bordering loessial scarps. A minor error in his Figure 8 is the bar scale by which the 60 miles of length measures 120 miles.

that Flint claimed. With this denial, the writers agree.

An outstanding infelicity in Flint's concept of a Cheney-Palouse fill is the quantity of debris demanded before the scabland was made. An average thickness of 100 feet on the 800-square-mile tract north of Hooper would require a volume of 15 cubic miles. Subtracting for porosity and glacial contributions, this still remains a fantastically impossible quantity to derive from subjacent basalt before scabland was eroded.

When Upper Grand Coulee's cataract receded through the monoclinal uplift, the new channel head thus afforded was 800–900 feet lower than any other in the complex. All other scabland channels then ceased functioning, unless for a possible very brief late closure of Grand Coulee by glacial ice. Hence the removal of Flint's valley fills must have been completed before the opening of the low Grand Coulee route.

A dozen separate valleys east of Grand Coulee led glacial water to Quincy basin (Pl. 1). Their supposed fillings must have been carried off to that basin before Grand Coulee's capture of their rivers. If we grant that there was time enough for this, was there sufficient capacity in Quincy basin to contain their debris in addition to what the great Coulee itself contributed?

The profusion of granite boulders along Rocky Ford and Willow Springs channels in Quincy basin gravel indicate that they were in use before Upper Grand Coulee's receding cataract had reached the Columbia valley. Drumheller Channels were well deepened by Rocky Ford and Willow Springs discharge, and Quincy basin's initial storage capacity was correspondingly decreased. Yet there is clear evidence that Upper Crab Creek glacial river was still functioning. By Flint's reasoning, it was still discharging gravel into Quincy basin.

Neff estimates (from U.S.B.R. data) that a "bare minimum" of 11 cubic miles of rock was eroded in making Grand Coulee and associated distributary coulees, all of it carried into Quincy basin. He finds that, if this debris had remained in the basin, it would have made a fill three or four times as thick as the existing deposit of scabland gravel. Thus this basin's fill, even if restored to a pre-channeling plain, cannot possibly represent the total debris contributed alone from Grand Coulee and its scabland complex. Great quantities must have gone on out by way of Drumheller, Othello, Koontz, and Lower Crab Creek. Add what was yielded by erosion in these deeply bitten dischargeways. Consider that all are largely empty today. Note that Pasco basin, also largely empty, lies almost immediately downstream and its highest "remnant" of Flint's or Allison's fill, Priest Rapids bar, 860 feet A.T., forbids a gradient for contemporaneous erosion of virtually all these channels. Where has this debris of the hypothetical fill gone if not into, and through, Pasco basin? Lake Lewis did not concide with any scabland floods.

Repetition of Floods

(Bretz)

If floods of comparable volume repeatedly swept down the Cheney-Palouse tract, scabland making on the Palouse-Snake divide would be resumed each time along the pattern left by the preceding flood. This behavior would be changed only after Palouse Canyon's upper falls (now only 20 feet high near the head of the 200-foot canyon) had retreated across the divide and only if later floodings could be carried off through it and the old Washtucna Coulee route. Did more than one flood cross the summit of the divide?

One approach toward an answer lies in relations of Palouse Canyon to the bars south of Snake River at the junction. The top of the mid-canyon bar is 130 feet higher than the foot of lower Palouse Falls, only 5 or 6 miles distant, and 60 feet higher than the foot of extinct H U Ranch cataract. Were these "waterfall" canyons eroded *beneath* the deep torrent that made the bars, or are they products of a later flood?

Matthes (1948, p. 258–259) has described the intense, quasi-vertical, vortex suction in turbulent streams (his Type III) as kolk action, calling it "probably the most important macro-turbulence phenomenon in natural streams" and "the most powerful form of concentrated energy at work on stream beds." Once the unparalleled magnitude of scabland streams is realized, the kolk, which is essentially a subfluvial tornado in structure and in velocity relative to surrounding turbulence, can readily

be accepted as the lifting mechanism in excavation of the great rock basins.[37]

Matthes notes that the most effective plucking action of a kolk occurs on downstream-facing vertical faces where cavitation is at a maximum. Celilo Falls, at the head of the Dalles of the Columbia, is 20 feet high in low-river stages but disappears during floods when it becomes a "subfluvial cataract" with maximum kolk action on the same rock in which the many scabland falls were developed. Can this idea be extended to include the production of subfluvial erosional forms which, after the stream has vanished, look like abandoned waterfall cliffs, plunge pools, and recessional cataract gorges?

The lack of continuous marked channels leading across the Palouse-Snake divide to the several abandoned cataracts is comparable to the relations of cataracts on Trail Lake anticline's southeastern limb. It debars any reasonable interpretation of such cataracts made by individual streams across either divide. Development beneath a wide sheet of water is demanded, and this would require subfluvial origin by kolk action.

If H U Ranch and Palouse falls and their recessional gorges are not explicable by the kolk concept, then there must have been some marked change in conditions, particularly in the level of water or debris in Snake River Canyon, after bar building and before completion of scabland canyons, a change at least suggesting two episodes of flooding across the divide.

Another approach toward an answer is suggested by the much smaller proportions of the short channel leading to the H U Ranch cataract than those of the cataract itself before the notching of its lip. It is comparable to the minor channeling above the lip of Potholes cataract and suggests the same thing, a later and smaller flood.

The lowest brink of H U Ranch falls has the same altitude as the top of the shoulder bar although 7 miles farther up the glacial stream. Most of the rim of the Davin gorge is only 50 feet higher. Unless the cataract and gorge are of subfluvial origin, or adequate lowering of water or debris in Snake Canyon occurred during a given flood, this third approach indicates that two episodes of flooding are required for these relations, the shoulder bar dating back to an unnotched brink of the falls.

A fourth approach lies in scabland canyon relations in the vicinity of Davin Ranch (Fig. 19), west of the large group of residual loessial hills interrupting the summit scabland of the divide. Several square miles of prescabland Palouse loess topography lies between Palouse Canyon on the east and the H U Ranch cataract and Davin gorge on the west, and extend south to the brink of Snake River canyon. A small scabland coulee crosses this residual tract, from northeast to southwest, and enters the gorge at the Davin Ranch. It heads in a group of irregular channels which are cascade chutes rather than cataract alcoves. The last mile of this coulee is wide and gravel-floored. It obviously is a modified, minor, preglacial tributary of the Snake, as also is the last 3 miles of the Davin gorge.

This tributary's gravel flat hangs, without any notching, about 140 feet above the adjacent gravel-covered floor of Davin gorge. This seems too great to ascribe entirely to excess depths of water in the main channel over that in the tributary. It seems obvious that the gorge floor was lowered after the tributary ceased to flow. Although water for both the tributary cascades and the main cataract had to cross the unchanneled divide summit at 1280 feet A.T. before entering either one, the channel to the cataract headed 3 miles farther upstream than did the cascade channel. Also three other cascade or cataract gorges, tributary to Palouse Canyon, head at this altitude within a mile of the H U Ranch channel head. All must have operated after Davin's tributary cascade chute ceased to function.

A fifth approach lies in two sets of glacial river-bottom records, separated by a vertical interval of 150 feet or more. A channel around the southern slope of the semi-isolated hill and its appended shoulder bar (Fig. 21) has a floor at 840 feet A.T., approximately the same altitude as the Davin gorge floor at the junction of its hanging cascade-chute tributary. Dis-

[37] Mr. Matthes permits us to quote from a letter that "as regards kolk action, the negative pressures involved would easily account for effective plucking, especially in the case of columnar basalt whose shrinkage cracks would facilitate plucking." He adds "I like your comparison of a kolk to a tornado".

tances to the Snake are 2 and 1½ miles, respectively, and the mouthings in Snake Canyon are only 2 miles apart. In contrast, the shoulder bar's flattish summit (at 1000 feet A.T.) is 160 feet above the channel to the south, and the cascade chute's gravel fill (between 960 and 1000 feet A.T.) is 120–160 feet higher than the Davin gorge floor. These relations strongly suggest that two floods crossed the Palouse-Snake divide, the earlier one making the shoulder bar and the cascade-chute fill, the later one finding lower levels (of ponded water or a debris fill) in Snake Canyon and consequently scouring out the channel south of the bar and lowering the floor of the gorge. The abundance of large boulders on the brink of the Davin terrace overlooking Snake River and the unfilled plunge-pool trench at the foot of the falls (Fig. 19) testify that H U Ranch cataract was operating at high efficiency to the very end of the second flood.

Cheney-Palouse Tract North of Palouse-Snake Divide

Maps: Benge, Haas, LaCrosse, Starbuck and Washtucna quadrangles; U.S.G.S.

Multiple Scarps

(Bretz and Smith)

A traverse from Sprague to Washtucna (Pl. 1) and a re-examination of the Willow Creek bar constitute all the study given to the tract north of the Palouse-Snake divide in 1952. The traverse was made to find multiple scarps in the bordering loess bluffs, features which Flint had stressed as "conspicuous" evidence for his episode of dissection following the aggradation. Bretz had no notes or memories regarding such things.

Four possible multiple scarps were seen in this 50-mile traverse. The lower member of each double scarp was a rounded form, not a sharply cut bench, appearing low on the slope and having a narrow tread. Time did not permit examination on foot, and comments here are only suggested alternatives to Flint's explanation. One suggestion is that low gravel bars deposited against the base of the loess scarps and subsequently scarped in turn would produce this effect of erosional terracing in the loess (Pl. 16, fig. 2). Another suggestion is that masked basalt ledges may be responsible.[38] Another is based on Culver's report (1937, p. 58) of probable Ringold remnants in the basal portions of "prow-pointed hills." Another, thought to be supported by other features of a particular case, is that they are remnants of the prescabland erosional topography, so situated that the upper, steep slope appears to belong to the scabland ensemble. A fifth suggestion is born of finding buried, calichified zones elsewhere in the Palouse loess. Such, being more resistant, could produce the effect of multiple scarps where subjected to lateral erosion. A sixth alternative comes from locating Flint's most convincing double scarp (his Pl. 6, fig. 1). It is a mile back up a narrow nonscabland tributary of Washtucna Coulee and 250 feet higher than the coulee floor. Although loessial scarps facing the coulee along this stretch have bases a little higher than this, only back filling could have occurred here. At the entrance to Washtucna, this tributary has a heavy gravel deposit almost as high as the base of his double scarp. (*See* Haas geological map.) The feature he illustrated is, as he admitted, a record only of the local stream's work, which surely is postscabland.

However, double scarps could be produced by glacial-river strands which did not, in their first operation, cut down to a basalt floor and hence had opportunity in a later flooding to deepen in loess. This procedure would be favored if an adjacent larger strand had been deepened sufficiently during the first episode to give exceptional gradient to the smaller one in the second. Indeed, eight of the double scarps listed by Flint are closely associated with such small

[38] Bryan (1937) figured a loessial island in the Cheney-Palouse tract near the Ralston divide crossing; his figure A of Plate 1, shows 50 feet of basalt in the lower slopes of its scarp. Bretz studied that hill after Bryan had called his attention to it. To get 50 feet of basalt requires measurement from the bottom of a closely associated scabland channel, and only the northern end of the hill has any basalt in the scarp. A low, south-facing ledge of prescabland origin, traceable for a few miles laterally out across the scabland on either side, provides this basalt. Enough retreat of the ledge occurred under the scabland river's erosion to leave a foundation of basalt in the lower part of the scarp. Bryan notes masking of the basalt in the scarp "by slump of soil from above."

channels separating small loess islands, and only three face large channelways.

Although Allison (1941, p. 55) described no such topographic forms, he accepted Flint's statement regarding their common occurrence but derived the multiple scarps from local ice jams.

Meander Scars

(Bretz)

Loessial islands are strikingly streamlined with the glacial-river courses (Pl. 16, fig. 1), and most of them have sharp prows pointing up the gradient; obviously they have been made by glacial-stream erosion. But do these narrow, elongated, residual hills and the narrow, elongated channels separating some of them fit the pattern that would be left by a stream (or streams) as small as the Snake, meandering on "great areas 10 to 20 miles wide"? Where are the meander scars in the loessial scarps?

Flint did note four examples of "crescentic undercut faces and slipoff slopes" (1938, p. 475) in the scabland deposits, none of which remotely approaches the Moses Lake pattern, but he did not state, as he did for multiple scarps, that there are "scores" of cases. There is no comment in his paper indicating that the "scarps and benches streamcut in the 'Palouse soil' . . . , recording streams cutting laterally on various profiles" (p. 472) bear any meander scars. Topographic maps, aerial photographs, and field inspection have revealed none to the authors.

Flint's "leisurely streams with normal discharge" began their work of degradation in many places on a fill banked up high (90 feet in his fig. 4) against confining loessial slopes. Removal of such fill to expose, and in most places to scarp, the buried loess apparently involved no normal meandering. Indeed, he found places where the relatively coarse stream fill had been removed and pre-fill, *unscarped* slopes of the fine-textured loess "exhumed by rinsing rather than by the more drastic process of lateral planation" (p. 486). This process of "rinsing" is not quite convincing.

Eardley (1938) was impressed by the "magnitude and manifest rapidity" of Yukon River meandering. Reports of residents and ages of trees in abandoned channels led to his estimate of 1000–2000 years for a meander loop to shift across the 10-mile (average width) flood plain. He judged the Yukon to be degrading its valley floor in the process. Concomitant widening, where meanders undercut valley walls, is largely in frozen "Yukon silts", and the greatest width attained is 35 miles.

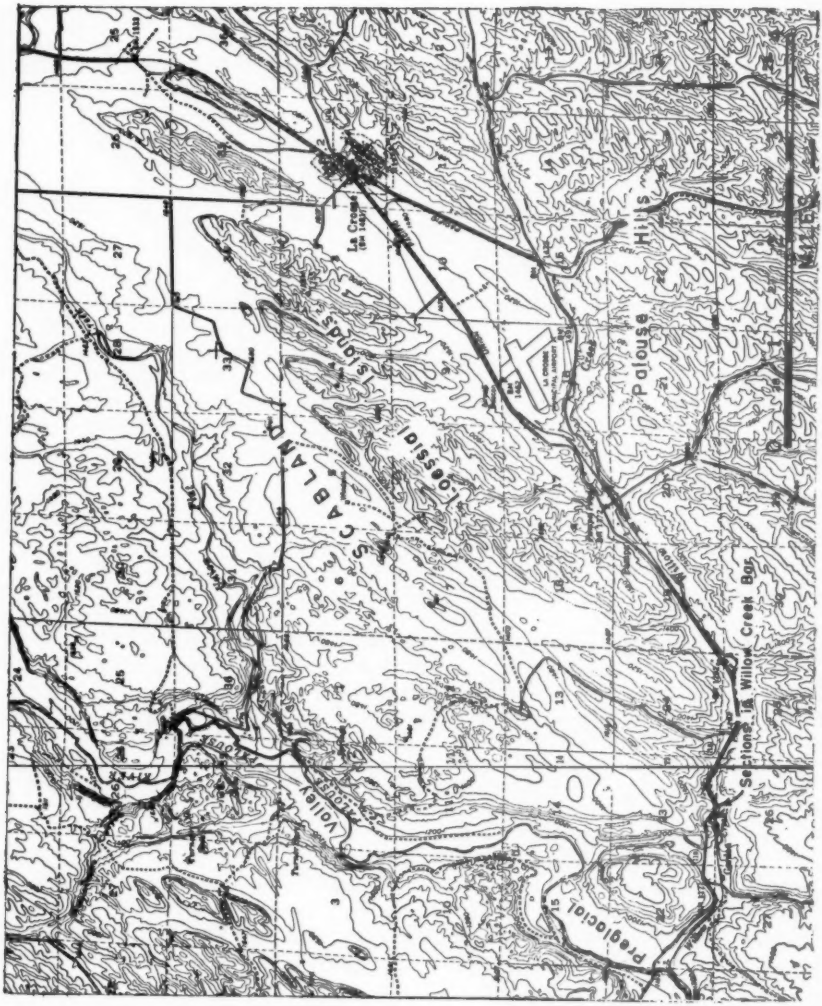
The loessial bluffs undercut by Flint's Cheney-Palouse glacial river may well have been similarly frozen, but his stream's load was much coarser than is the Yukon's, its gradient much greater, and volume of water much less. Any possible meanders therefore were smaller and less active. One wonders how long this scabland river's meanders required for removing the great fill of Flint's hypothesis—*i.e.*, how slowly the Lake Lewis dam wasted away. Valley glacier and landslide dams in historic time have had very brief lives compared with Eardley's estimate for one valley-wide meander sweep. Epeirogeny as first choice for the cause of the fill-and-cut sequence would have escaped this particular criticism.

We can accept heavily laden, braided, aggrading streams with gradients of 13 feet to the mile (Flint's figure) but not degrading rivers approximating the size of the Snake which meandered on gradients increasing to 20–25 feet per mile. Allison (1941, p. 56) called "impossible" Flint's concept for removing the fill he postulated. The writers agree. Furthermore, they believe that such streams could not have eroded the bedrock scablands, even with their final hydraulic gradient doubled.

Willow Creek Bar

(Bretz)

The new Benge and LaCrosse quadrangle maps show that Willow Creek flows from a typical, mature Palouse loess topography into a flat nearly a mile wide south of LaCrosse. The creek skirts this flat in a ravine about 50 feet deep and thence enters a narrow valley, which it leaves to traverse scabland to the Palouse. The creek still follows its prescabland course, but the marked valley widening south of LaCrosse is due to convergence there of six or seven lateral glacial channels from the north, separated by elongated loessial islands (Fig.


Palouse Hills
Pleistocene
Willow Creek Bar
Sections 13
Valley
Coulee
Coulee
Coulee

22). Three of the channels are bottomed on basalt at their heads and the broad flat is surfaced with basalt gravel.

The Willow Creek bar lies almost at the valley's entrance into scabland, in sections 23 and 26, T 15 N., R. 38 E. Bretz (1928a, p. 647–648) described it as a ridged gravel deposit once built across the creek valley and subsequently trenched through by a narrow, postglacial gulch. Flint (1938, p. 478)—finding (1) cut-and-fill recorded in lenses of silt included in this deposit, (2) fore-sets dipping both up and down stream, and (3) failure of some to dip parallel with the surface slopes—claimed that this valley barrier was "a remnant of a formerly continuous fill."

Allison (1941, p. 68–69) agreed with Bretz that the bar is constructional and wondered "how ordinary streams could have removed [Flint's] fill in the scabland west of the bar and yet have left the numerous Palouse islands and the seven channels across the divide west of LaCrosse except under conditions of ice blockade"—conditions which he did not specify. One unavoidable condition was that the ice blockade must have been 10 miles from end to end across the entire Cheney-Palouse tract (with one central interrupting loessial island) and about 400 feet thick in the deep preglacial Palouse River and Cow Creek valleys on either side of the island. Another limiting condition was that the supply of floating ice must have been abruptly terminated on completion of the dam, or the dam must have promptly failed; otherwise the seven narrow channels could not have escaped similar blockading.

As shown by fore-sets in highway, railroad, and creek trenches, the Willow Creek bar is composite, composed of basalt gravel brought upvalley from the west, and sand and light-brown silt brought downvalley from the east. A marked disconformity shows that there were two episodes of deposition separated by an interval of erosion.

The form of the earlier deposit no longer exists. Its exposed gravel is largely horizontally bedded but has eastward-dipping fore-sets locally. It dammed Willow Creek, causing later deposition of silt on the upstream (east) side, lenses of which tongue out westward into gravel rehandled by the creek during the erosional interval. The creek subsequently cut a ravine through the deposit and, as shown in the highway section, made a steep, east-facing bluff.

During the second depositional episode, scabland water from Palouse River valley swept up lower Willow Creek, eroded the west slope of the first deposit, tossed gravel over the top and down the face of the stream-cut bluff to build the conspicuous, long, east-dipping fore-sets that occupy the whole vertical range of the east half of the highway cut. The bar thus reconstructed a barrier to the creek, one that the stream has again succeeded in breaching. The earlier deposit may also have been a bar, or it may record a prescabland Palouse-Washtucna valley train, here built back in deltaic fashion into a tributary valley's mouth.

If the bar form were only a remnant of a former continuous fill in Willow Creek valley, why is it lingering in this narrow place[39] where six or seven glacial channels from the north discharged across it and where the drop today is 80 feet along the creek, and 200 feet from the bar top, to the Palouse 1½ miles away? It should long since have disappeared as did most of the glacial-river deposits under which Flint buried the whole, broad, flattish Cheney-Palouse tract. Its location is the last place to expect a lingering remnant.

FIGURE 22.—LOWER WILLOW CREEK AND VICINITY
Showing loessial islands, glacial-river channels separating them, part of Cheney-Palouse scabland, and location of Willow Creek bar. Part of Benge and La Crosse topographic maps, U. S. Geol. Survey. Contour interval 40 feet.

SLACKWATER SILTS NEAR PALOUSE-SNAKE JUNCTION

(Bretz)

The fill-and-cut or ice-jam theory may appear to find support in the slackwater silts back in valleys tributary to the scabland and in the Snake and its tributaries upstream from Palouse River junction (Bretz, 1929; Lupher, 1944). These silts are similar to the Touchet beds, although commonly they record currents flowing up the tributaries. They do not occur near the

[39] At 50 feet above the creek, the ravine across the bar is a third to a quarter as wide as the creek valley for the next 4 miles upstream, and at 100 feet above, it is less than half as wide. Summit of the bar ridge is about 30 feet higher than the creek at the transection, and less than 1300 feet A.T. Flint's fill was nearly 450 feet thick here (1938, p. 478), reaching 1700 feet A.T. Thus more than 400 feet has been eroded away to leave a mile-wide valley bottom just upstream from this constricting ridge at right angles to the valley length.

torrent-made bars in the region of the Palouse-Snake junction, suggesting that they are contemporaneous. Bretz has no very definite interpretation to make, except that (1) they clearly are deposits from glacial water that was flowing back up the tributaries from the Cheney-Palouse scabland and Snake River canyon, (2) they record some rhythm or repetition of conditions, and (3) they are only mantles ranging through a maximum of 700–800 vertical feet on the slopes. They are not terraced remnants of a continuous fill up to their upper limits which are essentially those of floated erratics upstream in Snake River canyon, 1300 feet A.T. Some exposures in valley bottoms probably are in materials reworked by postglacial streams.

Fresh highway sections were available in 1952 in the bottom of lower Tucannon River valley, which enters Snake River just upstream from the mid-canyon bar. Here a massive silt without current bedding appears to be the basal member. It could be a turbidity-current deposit. Disconformably overlying it is a fine sand with scattered pebbles, some consisting of the massive silt, and with some thin gravel units. At the base of this member may be a breccia of the massive silt. Channel cut and fill is common along the disconformity. Fore-sets record an upvalley current. Coarser material lies above and intertongues with this member. This also contains upvalley fore-sets. Sorting is very good locally in this third member, and there are some silt lenses and pebbles of foreign rock.

Equally significant but somewhat different records of a surge from scabland channels back up preglacial tributaries with no other access to glacial water exist in every valley entering the Cheney-Palouse tract for 50 miles north of Snake River canyon (Bretz, 1929). The deposits constitute a mantle of loess-derived silt with abundant angular grains and granules of unweathered basalt, a scattering of pebbles of foreign rock, and some erratic boulders. Commonly this mantle is unsorted and unstratified except locally on valley floors. Where bedding exists, it commonly records upvalley currents from the scabland. The upper limit in any one tributary valley corresponds to the upper limit of glacial-river water at junction with the main dischargeway. The deposits are thickest and coarsest in the tributary debouchures to scabland, and any stratification in such places is prevailingly fore-set out of the scabland and up the tributary. In numerous places, they are mounded scabland gravel deposits with long delta fore-sets comparable to the bars already described.

The outstanding feature of this backwater silt is the record of repeated upvalley currents during deposition of as much as 40 feet of material. No record of downvalley currents, except where postglacial rehandling is probable, have been found. A bursting dam farther down the Snake or Columbia would inevitably make strong currents downvalley. The coarser character of the material near the valley mouths and its gradation, in the farthest upvalley deposits, into a fine silt without sand or granule gravel units seems to indicate that it is gradational with scabland sediments, but neither the fill-and-cut theory nor the ice-jam theory can account for the reverse currents, the mantle-like character of the deposits, or the succession of events recorded. Repeated upvalley surges from the scablands could leave such a record.

Perhaps two kinds of backwater episodes are recorded in these deposits, both with upvalley currents. The well-sorted, well-stratified valley-bottom deposits may be the consequence of meltwater incursions from normal valley-train building along the main scabland routes. Because, however, floods have all but destroyed any record of valley trains, survival of such valley-bottom sediments seems anomalous. Furthermore, their existence in Snake tributaries above the Palouse junction requires a Palouse canyon, an Allison ice dam, or a Flint fill in order to cross the divide.

Glacial Lakes Missoula and Coeur D'Alene

Maps: Greenacres, Medical Lake, Oakesdale, Packsaddle Mtn., Priest Lake, Rathdrum, Spokane 1911 and Spokane N E 1950 quadrangles: U.S.G.S.

(Bretz)

The Pleistocene history of the region is still only sketched, but the writers believe it reason-

able to postulate that Glacial Lake Missoula's history included several dammings and several burstings. Pardee (1942) has described giant current ripples, huge bar deposits, and greatly scoured salients along and north of the narrow Clark Fork Valley portion of the Lake Missoula basin, produced by "a sudden failure of the ice dam that blocked this valley" and the escape through it of a large volume of water impounded in the Rocky Mountain Trench. He calculated the discharge as nearly 9½ cubic miles per hour, and his title, "Unusual currents . . . " is a fine example of understatement. Continuous discharge at this rate would empty the lake in about 2 days. But Lake Missoula could not have been emptied across the plateau until Upper Grand Coulee's cataract recession was completed. Extensive, thick, varved silts and clays, with glaciated stones dropped from bergs, probably tell of long-continued occupation of the deeper portions by a much smaller glacial lake or lakes. The varved sediments in the Bitterroot Valley portion lie a few hundred feet higher than comparable deposits in the Rocky Mountain Trench, yet are farthest from any lobe of the Cordilleran ice-sheet. There are outwash gravels above silts and similar gravels beneath silts, and relations vary in different arms of the very irregularly outlined lake. Pardee's illuminating discoveries[40] are only the beginning of what eventually should be a series of detailed studies of glacial pondings in western Montana.

Bretz has examined the mountain summit at the extreme northern end of the Bitterroot Range where the Clark Fork and Pend Oreille valleys diverge southward, (Figs. 23, 24) and against which the glacial dam for Lake Missoula was built (Priest Lake quadrangle map, Idaho-Montana). There is no notch at 4200 feet where Davis (1921, p. 140) speculated on an outlet. The aneroid read 4800 feet for the lowest saddle seen on the ridge, a saddle that has no channel features.[41] No suggestion of glacial ice above 3550 feet was found in the northwest branch of Johnson Creek valley, the route followed to the summit. The highest record found, a granite boulder, may have been berg-floated. The summit of the cliffed northern end of the range is 2000 feet above Pend Oreille Lake surface only a mile away. The steepness of the ice front here makes feasible a concept of catastrophic emptying of the lake, once weakening of the dam began and hydraulic quarrying under a head of 2000 feet of water took over. Lake Missoula probably never had an outlet, other than channels across, leakages through, and failures of its glacial dam. Had the dam remained intact through any one glacial episode the lake certainly would have risen to the lowest saddle. Discharge there would have channeled it and made a striking cascade chute down the western slope.

The multiplicity and very immature development of Lake Missoula's shore lines argue against a long stillstand at any one level. Although the highest of the series is no stronger than any others, the universal faintness may be due in part to submergence as the basin became filled. If there were several fillings and emptyings, the weak and close-set shore lines may constitute a composite aggregate of later mountainside scorings interpolated among earlier ones, the vertical sequence not being one also of time as Eakin and Honkala (1952, p. 1361) conceived it to be.

Although the late Wisconsin (Polson) moraine in Flathead Valley may have been built subaerially when no Pend Oreille glacial dam existed, and the shore lines on this moraine inscribed later (Nobles, 1952), it does not follow that there had never been any earlier pondings in the Lake Missoula valleys. The Elmo moraine at the head of Dayton Bay, on the west side of Flathead Lake and contemporaneous with the Polson moraine, has a few faint Lake Missoula

[40] G. A. Thiel (1932, p. 452–458) has described giant current ripples made by an observed abrupt emptying of a Minnesota lake. The ripples were 30 to 50 feet apart. They showed size sorting; coarser fractions of the debris were left in the troughs as winnowing carried the finer gravel forward to the crest of the ripple. Fore-set bedding in the ripples conformable with the lee slope was not reported, although Pardee noted and figured such structure in his Lake Missoula giant ripples. This item of fore-set structure is lacking also in the argument of the present writers because no sections were found in the scabland ripples, and there was no opportunity to make any. Neither were lag-cobble collections seen in the troughs among the scabland ripples, perhaps because of inconspicuous display, perhaps because the rippled gravel was already well size-sorted.

[41] Packsaddle Mountain quadrangle map, U. S. Geol. Survey, published subsequently, shows a lower saddle farther south, but it is 4600 feet A.T., 400 feet higher than the lake's highest shore line.

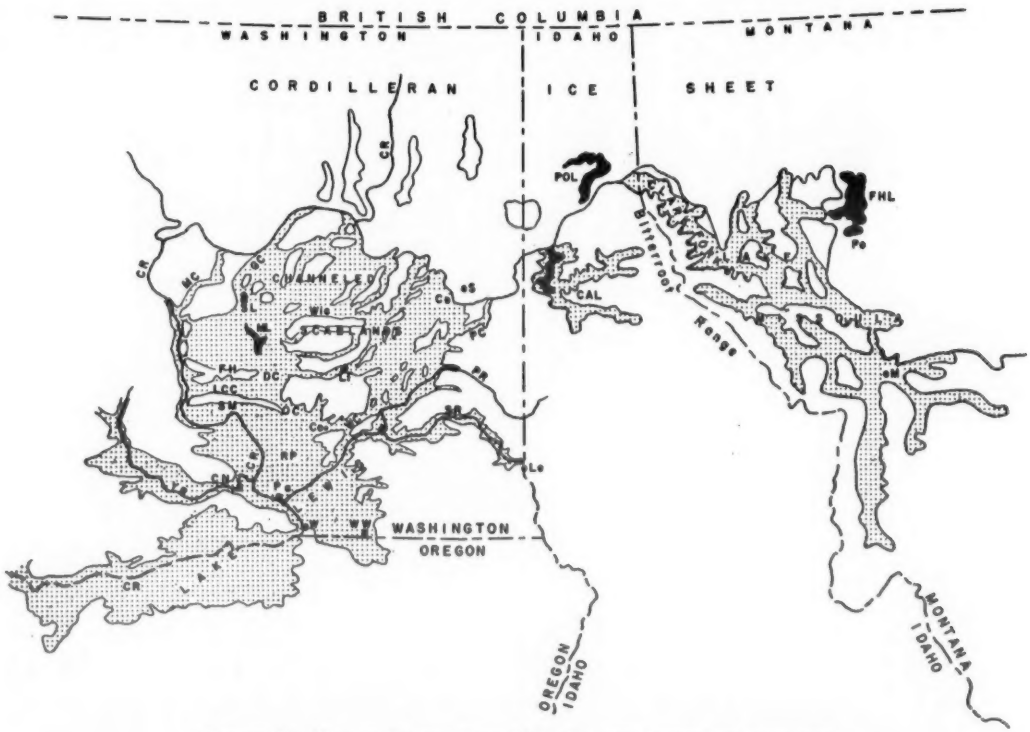


FIGURE 23.—MAP OF GLACIAL LAKE MISSOULA AND CHANNELED SCABLANDS

C—Cheney
CAL—Coeur d'Alene Lake
CN—Chandler Narrows
Co—Connell
CR—Columbia River
DC—Drumheller Channels
E—Elmo
FH—Frenchman Hills
FHL—Flathead Lake
GC—Grand Coulee
LCC—Lower Crab Creek
Le—Lewiston
M—Missoula
MC—Moses Coulee
ML—Moses Lake
OC—Othello Channels
Pa—Pasco
PC—Pine Creek
Po—Polson
POL—Pend Oreille Lake
PR—Palouse River
RP—Ringold Plain
S—Spokane
SL—Soap Lake
SM—Saddle Mtn.
SN—Snake River
W—Wallula
WC—Washtucna Coulee
WiC—Wilson Creek
WW—Walla Walla
YR—Yakima River

shore lines, whereas the valley slopes just outside the moraine are well scored with numerous fairly definite shore lines, certainly older than the moraine. Furthermore, outwash from the moraine occupies a trench cut in a valley floor of lake silt. There may have been no Lake Missoula in the southern part of Flathead Valley when the Polson and Elmo moraines and the Elmo outwash were deposited, but there was one before and one afterward.[42]

One critical region for the theory of a Lake Missoula catastrophic burst is the southwestern prong of the Purcell Trench. It extends from the site of the dam to the deep Spokane and Columbia valleys which skirt the northern edge of the plateau where overflow into the scabland channels occurred (Fig. 24). The trench, confined by highlands rising above the plateau level, contains remnants of basalt flows that were backed up into it. It also contains a great gravel deposit for its full length southwest

[42] W. C. Alden's long-awaited *Physiography and glacial geology of western Montana and adjacent areas* (U. S. Geol. Survey, Prof. Paper 231) was distributed just before the manuscript for the present study was submitted. Alden's (1953) comments on glacial lakes Missoula and Coeur d'Alene and the plateau scablands involve the following items.

(1) (p. 155) The 4200-foot saddle or col across the summit of the Bitterroots that Davis thought might be a Lake Missoula outlet was similarly interpreted by Alden. This idea is based only on study of the 1913 Priest Lake topographic map (scale 1:250,000 and contour interval 200 feet). Neither man could have made a field examination. Alden was "certain that the glacial ice dam was more than high enough to have held a lake up to the level of this col" (p. 143), but he obviously never had studied the Packsaddle Mountain quadrangle map.

(2) (p. 142) "There were probably at least three glacial invasions of the Purcell Trench".

(3) (p. 143) "It is probable that every time the Rathdrum [Pend Oreille] lobe of the Cordilleran ice advanced southward along the Purcell Trench, it crowded southeastward from Pend Oreille Lake basin into the valley of Clark Fork whose drainage became completely dammed . . . [forming a Lake Missoula]."

(4) (p. 155) "There seems to be no doubt that the mouth of the gorge of Clark Fork was dammed at least twice near the Idaho-Montana line. . . ."

(5) (p. 143) Glacial ice "must have crowded against the lower slopes of the mountains west of. . . . Lake Coeur d'Alene and must have overlain the 2500-feet basalt bench west of Coeur d'Alene [town]."

(6) Grand Coulee as a discharge route for water from Lake Missoula is mentioned (p. 146) under "Pre-Wisconsin Glaciation", its sill at the head noted as 1600 feet A.T. There is no comment on the time or conditions of Grand Coulee's origin; only the tacit assumption that the great gash then already existed at full depth. Under "Wisconsin Stages of Glaciation" Grand Coulee is again mentioned (p. 149) as part of the discharge-way for "another filling of glacial Lake Missoula", and it is also stated that glacial water may "have spilled over the part of the basalt plateau south and west of Spokane if the Grand Coulee was not open".

(7) (p. 155) "The relatively slight amount of wave work recorded along each of the . . . delicate shore lines of Lake Missoula would seem to indicate that no one stand of the lake was of long duration. The close spacing of the shorelines seems to indicate that the lake level fluctuated a number of times". Abrupt release of water from "lowering of the ice dam . . . might result in floods of great magnitude . . . even if Lake Missoula were only partially drained. Each may, *perhaps*, have been the origin of many violent floods that *are supposed to have* swept over the scablands. . . ." (italics ours). Thus Alden retreated slightly from his 1927 position.

(8) (p. 165) The Polson moraine "may have been deposited in the lake".

(9) Concerning Pardee's giant current ripples, Alden said (p. 96) of the "somewhat puzzling features" that "some of the numerous small ripples are beach ridges or bars . . . and some . . . may be due to moderate erosion of coalescent alluvial fans. . . ." No genesis for the large ripples, up to 50 feet high, was proposed. The most desperate efforts to escape the flood hypothesis can hardly dare use the above ideas for the current ripples on scabland bar summits.

(10) Alden admitted (pp. 96–97) that one of Pardee's channels (Rainbow Lake gap), contemporaneous with his current ripples, "looks as though it were at some time traversed by a river but its open eastern end now hangs hundreds of feet above the floor of the Camas Prairie Basin on the east and above the bottom lands bordering the Little Bitterroot . . . to the northwest". Instead of Pardee's explanation, Alden favored either a glacial river marginal to a possible Camas Prairie lobe (for which lobe existence he admitted that he could find no field evidence), or a normal river in Pliocene or early Pleistocene time, its drainage coming from the surface of a vanished basin fill of Tertiary sediments. But Pardee described the channel as "severely scoured, exposing the bedrock which remains bare. . . ." and as possessing "channel depressions, several of them holding ponds or marshes". Rainbow Lake, 1½ miles long, is the largest unfilled depression. A Pliocene or early Pleistocene age for the channel is very doubtful indeed.

(11) The massive "high eddy deposits" of Pardee, earlier called deltas by Campbell (1915, p. 142) and lateral moraines or morainal embankments by Davis (1921), Alden (p. 159) considered them to be gulch mouth fills or "lacustrine bars" made at different times during Lake Missoula's history. But harmonizing their bulkiness with Lake Missoula's "delicate" shorelines is impossible. The ridged valleyward margins of their summits and the undrained depressions farther back in the gulch mouths were explained as due to "settling and deepening" as "percolating water may have carried some of the finer material down into the interstices . . . [of the coarse material of the deposits]". Disposal of coarse contributions to these basins from subsequent gulch erosion was not discussed.

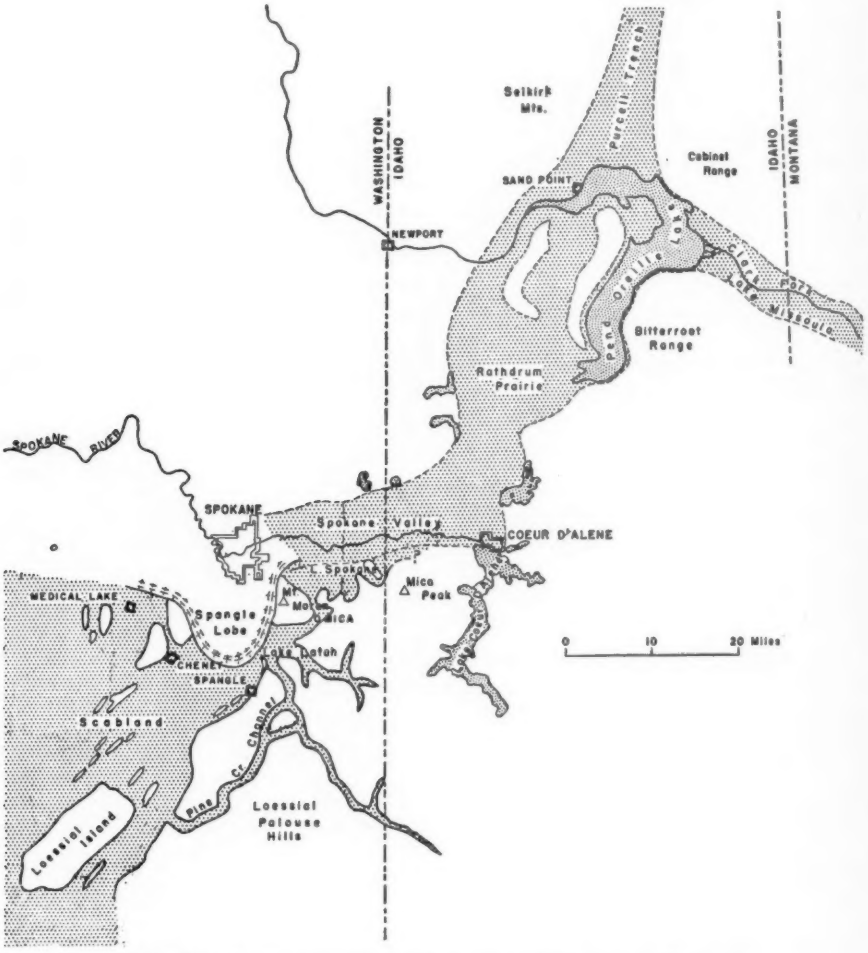


FIGURE 24.—REGION BETWEEN LAKE MISSOULA AND CHANNELED SCABLANDS

of the dam. This region has been examined only in reconnaissance fashion.

Lake Pend Oreille, at the site of the dam, is relatively shallow north of the entrance of the valley (Clark Fork) which contained Lake Missoula's westernmost portion, averaging less than 200 feet and containing several islands. South of this junction, the basin of the lake deepens abruptly to reach more than 900 feet in depth for nearly 15 miles along the length and more than 1140 feet for more than 5 miles. (*See* Packsaddle Mtn. quadrangle map.) Its maximum depth is reported to be 1800 feet. The lake surface is 2051 feet A.T. (2048 in 1946), and the bottom is therefore 900 feet or less A.T. in the southern part. This is 1000 feet lower than the brink of Spokane Falls nearly 50 miles downvalley to the southwest and is actually lower than the bedrock foundation of Grand Coulee dam more than 60 miles still farther west.

This deep part of the lake, constituting two-thirds of its total length, closely margins the western side of the northern tip of Bitterroot Range, one buttress of the Lake Missoula dam. A clean-cut mountain wall rises in 2 miles from

lake bottom to average altitudes of 4900 feet A.T. in Green Monarch Ridge. The lowest saddle in this ridge is 4600 feet A.T. (Packsaddle Mtn. quadrangle map). This deep basin can hardly be anything but an erosional product of its glacial experiences, its southern part now deeply filled with gravel which rises steeply 400 feet above the water, its slopes kettly and mounded from deposition against glacial ice (Pend Oreille lobe) then occupying the lake basin. The fill obviously is younger than the trough.

Between the site of Lake Missoula's dam and the plateau scablands, the large basin of Lake Coeur d' Alene (Fig. 24) extends southward from the Pend Oreille-Spokane valley, the southwestern prong of the Purcell Trench. Striated erratic boulders at many places around this lake up to 2700 feet A.T. obviously were berg-carried to position (Anderson, 1927). But field examination of the southern and western slopes enclosing Lake Coeur d' Alene shows that it never had an outlet directly to the plateau. Its highest level is to be associated with upper limits of the scabland channel heads west of Spokane, approximately 2500 feet A.T. The lake, at 2700 feet, had at least four times its present area of 60 square miles and was 575 feet deeper. It may have served as a great surge basin.

Glacial ice has apparently reached southward down this southwestern prong of the trench as far as the drift dam at the north end of Lake Coeur d' Alene 23 miles distant from Pend Oreille Lake. This portion, now deeply gravel-filled, is streamless Rathdrum Prairie. At Lake Coeur d' Alene the course of the trench turns more directly westward (the Spokane Valley) and opens to the basalt plateau at the city of Spokane. Here another lobe (Spangle) crossed the mouth of this trench, reaching some 15 miles south of the city, out on the plateau. Probably the Spokane Valley part of the trench was also ice-filled at this time.

At Spokane superposition of the river on a remnant of the basalt tongue that originally backed up the trench has made the Falls and protected the gravel fill upvalley from deep dissection. The fill does not, however, have a simple profile. In the Rathdrum Prairie section it is traversed by an empty channel 100 feet or more in depth and a mile wide, leading from Pend Oreille Lake as far southward as Lake Coeur d' Alene where Spokane River, the lake's outlet, enters it. In the Spokane Valley section, which extends thence westward to the Falls, terracelike deposits of stream gravel, fore-set away from the main valley, block the mouths of tributaries, making swamps or lakes in them. In the Rathdrum Prairie section, comparable ponding of tributary mouths has also made lakes (Coeur d' Alene being one), but the blockades are in part morainic and they stand higher than the Prairie surface.

Flint (1936, Pl. 6) considered that the inner valley followed by Spokane River has been eroded in an original fill which, still intact in Rathdrum, remains in Spokane Valley only in the gravel blockades to tributary mouths. This erosion entrenched the early fill at Spokane Falls about 150 feet and at the junction of Spokane and Columbia rivers, 45 miles farther downstream, as much as 200 feet. He also considered that the upper profile never extended farther down the Columbia but ended at or a little below 2000 feet A.T. because a lobe (Columbia) of Cordilleran ice blocked further transportation. How the meltwater passed the blockade and where it escaped were not specified. The lower surface, he believed, continued down the Columbia from this junction as a silt and sand deposit almost without gradient. The ponded water was discharged through Grand Coulee (after Steamboat Falls had receded through the monoclinal uplift) while the Okanogan lobe blocked Columbia valley farther west. Flint considered the Rathdrum Prairie portion of the Purcell Trench fill as the still-intact surface of the original fill.

Although it seems more probable that the lower profile in Spokane Valley is continued upvalley as the empty Rathdrum Prairie channel (not noted by Flint) than that the two profiles converge eastward to become one near the Idaho-Washington line, as he thought, all altitudes and characteristics of the deposit show that the entire sequence is postscabland, except for late functioning of Grand Coulee.

There are still higher stream deposits, however, lying in and marginal to the scabland channel heads a little west of Spokane; their altitudes are as much as 500 feet above the

upper profile in the mouth of the trench. If the trench east of Spokane contains correlative deposits, they must be the isolated outwash patches shown on Flint's Plate 6 (1936), 150–350 feet above his upper profile. The highest of these fall within the altitude range of the near-by scabland channel heads. The trench between the Lake Missoula dam and Spokane carries no other known deposits that might record glacial water on its way to any scabland channel head except Grand Coulee.

Although all the Cordilleran lobes which crossed to the plateau east of the Okanogan Valley must have stood somewhat short of maximum extent when the scabland channels functioned, the deep canyons of the Spokane and Columbia must still have been buried under glacial ice. It seems entirely probable that the Rathdrum-Spokane valley then also contained glacial ice and therefore that the flood discharge traversing it left no identifiable valley-bottom deposits.

There remains the possibility of *erosional* records, lying higher on the slopes. Five spurs on the southern slope of Spokane Valley and one saddle across the granitic hills separating that valley from the plateau to the south appear to be significant. One spur is a basalt bench just west of the north end of Lake Coeur d' Alene, at an altitude of 2400–2500 feet A.T. Its surface, 2 miles long and nearly a mile wide, has typical scabland topography of knobs and buttes and ridges separated by anastomosing channels elongated parallel to the higher valley wall, their total relief about 50 feet. That it is not a glacially plucked surface is clear from the character of glaciated surfaces on two other basalt benches much nearer Pend Oreille Lake, and in many places along the northern margin of the plateau. Glacial ice smoothed the Columbia basalt; it rarely plucked except where forced to climb north-facing slopes. This local scabland, 200–300 feet lower than the highest drifted erratics in near-by Lake Coeur d' Alene valley but equally higher than Rathdrum Prairie, appears to demand glacial water escaping westward between ice on the north and the valley wall on the south.

Glacial debris on this bench (see footnote 42, item 5) is probably contemporaneous with that in the lake basin's dam 2 or 3 miles farther east, and therefore belongs to the Rathdrum Prairie glacial drift of postscabland (except Grand Coulee), valley-fill age.

Four northward-projecting granitic spurs between the State line and Spokane possess scoured and channeled notches close to the higher, steeper valley slopes. Floors of the notches range from 2200 to 2480 feet A.T. The distal ends of the spurs are severely scrubbed whereas higher slopes are not. The spur nearest Spokane, a ridge that locally narrows the valley to about half, is just south of Dishman (Spokane N.E. quad.). Its terminal mile lies between 2000 and 2500 feet A.T. and is mostly bare and fairly fresh rock with a very irregularly rugose surface. South of it rises a steep cliff. Above that, the ridge has a cover of soil with protruding ledges of deeply weathered gneissic and granitic rock. The bare-rock part of the spur carries several transverse notches, some nearly 75 feet deep and containing marshes. As with the basalt ledge at the north end of Lake Coeur d' Alene, these features are ascribed to glacial water flowing westward between an ice mass in the valley and the rocky mountain side margining it, and escaping across the scablands. No argument for great volume can be advanced for these channeled spurs, but neither can it be denied.

The saddle across the range of old hills protruding through the basalt south of Spokane Valley is a broad sag containing two col notches a dozen miles southeast of Spokane, between Mt. Moran (3700) and Mica Peak (5205) (Fig. 24). The railroad station of Mica marks one col, from which drainage goes north into Chester Creek and southwest into California Creek (Greenacres quadrangle map). There is some channeling by glacial water in the deeply decomposed granitic rock at the cols, but obviously no marked current has ever traversed the saddle from Spokane Valley on the north to the plateau to the south. Erratic material is common up to 2550 feet A.T., but none was found higher.

Flint (1936; 1937) mapped the Mica saddle and most of the drainage area of California Creek as "drift-covered"—*i.e.*, glaciated—with Mt. Moran as a nunatak. It seems much more probable that the scattered erratic boulders and patches of till on which he based his inter-

FIGURE 25.—SCABLAND CHANNEL HEADS NORTHWEST OF CHENEY

The hills are of granitic and metamorphosed sedimentary rock projecting through the basalt of the plateau. Fish Lake Channel, used by three railroads, has the only basin eroded in basalt and the lake surface is 200 feet or more below the channel floor to the southwest, also on basalt. Thus there must have been fairly deep prebasalt notches across this ancient divide, much of their depths unobliterated by the lava flows and waiting for the glacial water that sharpened and channeled them. Glacial ice has never been over these hills, but it has covered the plateau to the north and has been in the channel heads since the scablands were made. Part of Medical Lake topographic map, U. S. Geol. Survey. Contour interval 25 feet.

pretation were berg-carried to position. They fall far short of a "cover" for California Creek drainage, and their upper limit is nowhere known to be above 2550.

The ponded water for this flotation was named Lake Latah (Bretz, 1923, p. 582) under the impression that it was dammed by the Spangle lobe at maximum extent. It has long since been reinterpreted as a part of an elongated body of water just north of the plateau divide which supplied all the scabland channel heads. The "outlet" for the "lake" was the easternmost channel of the scabland complex (Oakesdale quadrangle) (Fig. 24; Pl. 1), discharging southward from the flooded valley of north-flowing Latah Creek by at least three subparallel and subequal routes across the plateau divide to southwest-flowing Pine Creek and leaving a well-marked glacial river course (Fig. 24) thence for 20 miles down Pine Creek to its entrance into the Cheney-Palouse scabland tract. The combination of features from Mica to the mouth of Pine Creek channel may be interpreted as the record of an early flooding out of Spokane Valley before much deepening of the Cheney-Palouse channel heads had occurred.

The four heads northwest of Cheney (Fig. 25) contain nine named lakes in rock basins and a multitude of rock-basined swamps and smaller lakes. Five of these lakes lie essentially on the plateau divide. Altitudes of channel-head floors range from 2300 to 2424 feet A.T. With other channel-head deepening, later floods may not have risen to the bottom of the Mica col, 2475 feet.

Nevertheless, a marginal meltwater lake must have existed back of the plateau divide, and the abrupt arrival of a Lake Missoula flood must have produced some gradient to its surface. Upper limits of the "Lake Latah" part should, therefore, theoretically be higher than those farther west. Interpretation of Mica col remains uncertain.

Evidence submitted in this paper for great volume of glacial water discharging across the scabland-scarred part of the Columbia Plateau debars any view that it came from excessive melting because of climatic amelioration. Despite the present lack of convincing evidence for torrential discharge between the Missoula dam site and the scabland channel heads, there is no evidence to contradict the hypothesis that enormous volumes of glacially derived water were abruptly discharged across this part of the plateau. Acceptance of Lake Missoula as the source involves no further assumptions to defend the idea.

Tempting speculations arise from the concept of a Lake Missoula dam burst. They include:

(1) abundance of floating ice riven by the dam's collapse;

(2) magnitude of individual bergs (Bretz, 1930b, p. 410);

(3) character of turbulence (excluding the kolk), as conditioned by local channel volume and load and by inherited gradient and valley shape;

(4) varying velocity of advancing front along different channelways, as conditioned by differing channel lengths, proportions and gradients;

(5) differing time of arrival of flood fronts and crests at various convergences;

(6) lag in initiation of channels with higher original sills at heads;

(7) differential yielding of various basalt flows as conditioned by both original and superimposed structures and by varying volumes and gradients of individual streams;

(8) possibility that waning stages of a flood made records suggesting successive floods;

(9) possibility that subglacial leakage from Lake Missoula made the excessive depth of Pend Oreille Lake.

The writers have endeavored to avoid *a priori* speculations as much as possible. Almost every argument for great volume starts from adequate field evidence. The cataract cliffs, alcoves, and plunge pools, and the huge rock basins on preglacial divide summits exist. The enormous, stream-lined piles of fore-set gravel on channel floors are a verity. The giant current ripples are, we believe, unchallengeable. The upvalley currents in tributaries cannot be denied from the record. The scarped loessial islands allow no interpretation other than ours. Furthermore, there was a Glacial Lake Missoula, 90 to 175 miles distant from, and 1700 feet above, initial scabland channel heads, and it was emptied by continuous valley routes directly toward those channel heads.

Summary

(Bretz)

Bretz's Flood Hypothesis (1923–1932)

The greatly scarred linear tracts of denuded basalt interlaced in a complicated pattern and elongated with the dip slope of the otherwise loess-covered plateau of eastern Washington have been made by glacial floods entering a maturely developed normal drainage pattern. The adjective "channeled," first proposed by Bretz (1932b), and used in a dozen later papers implies that an enormous volume of water transformed the invaded valleys into gigantic river channels. Such great quantities could continue to arrive only for a brief time, and a large volume of ponded water, capable of being abruptly released, is needed for this interpretation. Lake Missoula's volume of 500 cubic miles (Pardee) back of a glacial dam with one buttress most vulnerably situated for a catastrophic failure is argued as the source. Even then, repetition of the cataclysmic flooding seems indicated by the amount of work done by the short-lived glacial rivers. This paper presents new field evidence believed to be consonant only with the interpretation of repeated floods. Its leading theme is that the associated mounded gravel deposits have constructional forms, are gigantic river bars.

But the demand for unprecedented volume has seemed outrageous to conservative geologists, and other hypotheses have been advanced to explain the extraordinary erosional and depositional features of the Washington scabland. Two of them possess sufficient apparent applicability to have secured adherents

Allison's Ice-Jam Hypothesis (1933)

Allison (1933) argued that the scabland phenomena resulted from a damming in the Columbia River gorge across the Cascade Range, more than 100 miles downstream from the plateau scabland. His dam was initiated by a landslide and composed of berg ice and river ice. Once started, it grew upstream by successive increments, eventually ponding the deep major valleys that bound the plateau and allowing glacial stream water of normal volume to cross low places in divides of the plateau's preglacial drainage system. The complexity of the scabland pattern required many other ice jams to provide for divide crossings farther back up the plateau slope. Water bypassing these ice jams in "run-around" lateral channels gashed into the blocked valleys' slopes to make high-lying scabland and remodeled older terraces to form "perched" bars.

Allison admitted that his Cascade Gorge dam must have been 900–1100 feet high but did not point out any field evidence for its precise location nor for its great "run-around" bypass channels in the gorge walls. Nor has he dealt adequately with the resultant flood down the Columbia (1932) below the dam when that great obstruction finally failed. He never located any smaller dams back on the plateau although he wrote about the excellent records left by stranded bergs in the widest ponding of glacial water in the region, the Pasco basin.

Allison's greatest dam lengthened upstream along the Columbia for 100 miles or more above its inception in the Cascade gorge. It thickened as it lengthened, and many diverted lateral strands of the glacial Columbia cut notches and channels across interfluves between entering tributary valleys and made gravel and rubble "dumps", "deltas", "fans", and "bars" in water ponded in these valleys. Several may occur at different altitudes in one locality, but, taken as a whole, they fall into a general descending profile for this valley stretch (Bretz, 1925). Eventually this ice jam was so thick that the Columbia became a series of ponds in these tributary valleys, standing at about 1100 feet A.T. for the entire distance (Allison, 1933, p. 722). No diversion channels were made at this altitude; only pebbly silts and berg-rafted erratic boulders attest the maximum flooding. How the glacially swollen river traversed this 100 miles of individual lateral pondings on no gradient is unexplained. How the diversion channels and gravel deposits near the main valley escaped modification while the dam was growing to full thickness is a puzzle for the deeper part of this long stretch of Lake Lewis was then almost solidly full of ice. The high-lying silts and rafted boulders may not date from the episode of diversion channels and gravel deposits. Leaving out the 1100-foot records, Allison's picture for this 100 miles of Columbia Valley is appealing but lacks any

positive records of the ice blockade itself or of a downvalley rush of water when it failed.

Search was made at Washtucna Coulee, Esquatzel Coulee, Lower Crab Creek, Upper Crab Creek, and Palouse Canyon, where divide crossings or lateral canyons seemed to suggest detours caused by ice jams in the glacial streams. No favorable constrictions and no records of stranded bergs were found. Instead, the upper limits of the glacial rivers, recorded in steepened slopes of the bordering loess, possess uniformly descending profiles along such glacial river courses.

Local ice jams may have occurred in the making of the scabland but no evidence for any has been found. Allison implied that downvalley floods from the inevitable bursting of such dams made the great mounded scabland gravel deposits (remodeled older terraces) which he recognized had largely constructional forms.

Allison also theorized on a deep fluvial fill in the capacious Pasco basin. This basin was undoubtedly occupied by glacial water up to the limits specified, but the only deposits which might conceivably record his detrital fill are huge, elongated mounds, 100 feet or more above the bottoms of closely associated depressions, both forms streamlined like the bars he identified in Snake Canyon, but lying out in the middle of a river valley more than 15 miles wide. The largest depression contains a closed basin $\frac{4}{10}$ mile long and more than 50 feet deep. He found only one terracelike gravel deposit in the basin to fit his theory.

Flint's Fill Hypothesis (1938)

Flint (1938) proposed that the many divide crossings and the abundant scarps in loess-covered tracts bordering the scablands had resulted from deposition in the valleys invaded, a common procedure of laden glacial streams but continued here until low divides were crossed and the plexus was outlined. Raising of base level to produce this deposition was presumably because of a dam in the Columbia gorge, formed probably by a great landside or a local valley glacier pushing down into the gorge. This dam was as inferential as Allison's. Flint's glacial streams across the plateau were shallow and "only small pieces of glacier ice" ever floated down them. During wastage of the dam, the glacial streams gradually removed their fills to leave remnants (15 per cent of original deposits—Flint) which he called terraces modified by creep and slopewash; "in various states of [subsequent] dissection." Every high-lying minor unit of the plexus was as well cleaned out as the larger, deeper ones. Wide-swinging meanders were a feature of the dissection on the broader (10–20-mile) tracts. The rugged basalt scabland, including the numerous and spectacular rock basins, was made largely during this removal.

This Columbia gorge dam must have wasted away very slowly, for the nearly complete removal of the fill in scabland routes was all done with moderate quantities of glacial water while also the valley bottom basalt was being scarified (by meandering streams!) into the butte-and-basin topography characterizing scabland.

Flint's interpretation of the gravel deposits associated with the scabland routes as "non-paired, stream-cut terraces", cannot be accepted. Many of the "remnants" are ungullied hills of gravel standing out in mid-valley, possessing stream-lined forms, having associated closed fosse depressions between them and valley walls, containing bouldery accumulations for the transportation of which his "leisurely" streams never could be held competent, displaying internal structures of long fore-sets parallel to the slope for the entire height of the deposit, and carrying gigantic current ripples on their summits and channel-side slopes. Allison, in a later paper (1941), recognized these gravel piles as subfluvially shaped individual forms made in favored places.

There are divide crossings inexplicable by Flint's rising fill (Drumheller, Othello, Devils Canyon, Palouse Canyon); there were contemporaneously functioning spillways out of basins (three cataracts and Drumheller from Quincy basin, Dry and Long Lake coulees out of Bacon syncline) which, even with adequate fills, required far greater volume of water than Flint allowed.

Flint reported that many loessial scarps margining the scabland river routes were "multiple" and argued that the benched slopes

recorded stages in removing the fill, overlooking any alternative possibilities.

The extraordinary pattern of islandlike, scarp-margined, prow-pointed residual hills of loess isolated in wide areas of scabland and cut through by narrow, subparallel accessory glacial river channels was to Flint the consequence of his widespread aggradation. But in the cleaning out of all the filled valleys by lateral planation, no meander scars were ever left in the loessial scarps.[43] Furthermore, Flint's mechanism "breaks down utterly" (Allison, 1941, p. 56) when catharsis of all the high-lying minor divergent strands is required of it.

Flint's arrow went widest of its mark at the place where Allison specified an ice jam in Snake Canyon at the junction of the Cheney-Palouse scabland river with the larger valley. For this district, Flint listed several dry falls and rock basins, stressing their small dimensions (too small to be shown by the 40 ft. contour intervals of the subsequently published topographic maps) as typical. Unfortunately, he missed seeing dry falls 100 to nearly 300 feet high, empty recessional gorges with closed depressions in them more than 100 feet deep, and rock basins equally deep on the rocky summit of the preglacial divide where a width of 9 miles was swept clean of its loess cover and 9/10 of its 80-square mile area has no remnants of his fill.

Flint hypothesized a debris fill completely blocking Snake Canyon just beyond this ravaged divide summit. To make a lake (Riparia) to account for high-lying, bergrafted erratic boulders farther up the Snake, the fill had to be 800 feet thick. The outlet of the lake had to be around the southern low periphery of the fill. In the subsequent dissection of this dam (during the slow wastage of his Columbia gorge dam), Snake River unaccountably abandoned the low marginal bypass route, where it should have become entrenched, and returned to its buried preglacial gorge. The fill is almost all gone, and the exhumed, rugged basalt topography forbids any concept of a lateral shifting to bring the Snake back.

[43] Before one commits himself as skeptical of the interpretation of the present study, he should examine the Benge, Haas, LaCrosse and Starbuck quadrangle maps. A topography probably without close approach to parallelism in his experience with maps or terrain is depicted on them. Some of the "islands" (minor prescabland interfluves) have had one side so largely removed that little valleys draining to the opposite side have been beheaded.

Character of New Evidence (1952)

The new evidence for torrential discharge comes from detailed topographic maps (10-foot and 2-foot contour intervals) by the U. S. Bureau of Reclamation; from new topographic maps by the U. S. Geological Survey; from 175 miles of main irrigation canals excavated, in large part, in scabland gravel deposits; from dozens of large, recently opened gravel pits; and from aerial photographs not available during any earlier study.

The aerial photographs reveal giant current ripples on mounded gravel deposits interpreted as gigantic river bars, essentially unmodified subfluvial constructional forms.

Long, deltalike fore-set bedding parallels the slopes and extends throughout the body of many of these bars. The large quantity of transported boulders in some bars and their huge size are impossible products of Flint's streams.

Records of repeated debaclelike floods have been found well distributed throughout the whole scabland complex. The western part of the scabland (Grand Coulee, Quincy basin, adjacent Columbia valley) has had at least seven floods, five down Grand Coulee and two affecting the Columbia valley alone. These are accounted for by repeated burstings (and reconstructions) of the glacial ice dam of Lake Missoula in the mountains of western Montana. Altitudes and topographic relations between the lake and the plateau are in consonance with the idea. Striking records that the lake was abruptly emptied (giant current ripples, severely scoured salients, huge barlike deposits, Pardee 1942) strongly support that interpretation.

Flood Correlations

(Bretz)

Correlations of all the postulated floods are not yet feasible. The latest flood across Quincy basin came from Grand Coulee alone, only after that deep canyon had been cut com-

pletely across the plateau's high northern edge and only when the Okanogan lobe, holding an advanced position across the Columbia valley, prevented escape of the glacial water around the northwestern part of the plateau. This event is generally considered to be of late Wisconsin age. At that time only deeper Drumheller and Othello channels and Lower Crab Creek operated downstream from Grand Coulee. Backfilling in the mouth of Upper Crab Creek and perhaps in lower Washtucna Coulee was contemporaneous with it.

Bretz has argued that all channels in the scabland complex functioned contemporaneously until Grand Coulee's great gash was completed. With this interpretation, Flint has agreed (1938, p. 46). But the new evidence of deepening of some channels during a succession of floods renders that concept invalid for later occupations. The idea is not yet to be dismissed for the earliest recorded flooding. It is based on altitudes of upper flood limits at the channel heads on the high northern edge of the tilted plateau. About a dozen definite channel heads are distributed along about 85 miles of this divide, and the upper-limit altitudes (highest scabland, loessial scarp bases, and ice-rafted erratics floated back in minor valleys in the loess) lie between 2450 and 2550 feet. A.T. (Bretz, 1928b, Pl. 5). Although these are not precise markers, there is no cumulative decrease in their altitudes either east or west. Part of this range may well be ascribed to aneroid errors.

The conclusion that the first recorded flooding used all channels can be avoided if the sites of the western channel heads were originally sufficiently lower than the eastern ones to take all the first flood water and (2) in some subsequent flooding, either these heads were closed by an advanced edge of the Cordilleran ice sheet so that higher eastern channels were initiated or eastward down-tilting had occurred since the preceding flood. This sequence can be reversed by blocking the lower western channel heads at the time of an early flood with an expanded Columbia lobe so that only the higher Cheney-Palouse route operated and then in later floods which found a shrunken Columbia lobe, bypassing that eastern route. Postscabland east-west tilting would again be necessary to bring all channel-head upper limits into the same horizontal plane. An eastward down-tilting, however, would remove the present difficulty with altitudes at Mica.

The channeled summit of Steamboat Rock in upper Grand Coulee carries glacial striae and a moraine ridge, both clearly later than the channeling (Bretz, 1932, p. 35). A silt-covered bouldery ridge on the coulee floor close to the cliffs of Steamboat is perhaps morainic. The eastern margin of the Okanogan lobe apparently has invaded the coulee head since cataract recession passed the Rock. If this completely closed the coulee, it must have caused renewed discharge of flood or normal meltwater through the deeper of the eastern scabland channels. Such an episode must have been brief, for very little retreat of the Okanogan lobe would again open Grand Coulee. A flood down Columbia valley west of the scablands followed the last Grand Coulee discharge, its record indisputable in West Bar's giant current ripples and the Beverly bar. Bursting of the Okanogan lobe's dam (Waters, 1933, p. 617), perhaps as a consequence of another Lake Missoula dam burst, may well have supplied the water for this latest glacial river flooding of the Columbia Plateau.

The writers do not believe that successive Lake Missoula emptyings are to be correlated strictly with successive Pleistocene glacial episodes. Dam failures probably came when increasing hydrostatic pressure in front of the Pend Oreille lobe coincided with weakening of the ice front. Decreased alimentation of the lobe would accompany sub-periodic climatic warmings when wastage would also be excessive. Repeated failures[44] may have occurred during any one glacial cycle. This would be comparable to recorded modern glacial lake bursting in Alaska, British Columbia, the Himalayas, the Alps and elsewhere (Kerr, 1934; Klebelsberg, 1948; Stone, unpublished).

The steep scarps in weak loess can hardly be survivals from early Pleistocene glacial floods. If, therefore, they were made in relatively late Pleistocene time, either there were no earlier Lake Missoula pondings and outbursts or the

[44] Pardee's giant ripples (1942) record a lake surface at least 3450 feet A.T. when the dam burst.

scarps record a late flood greater than any earlier ones.

It is significant that the closed basins of the scabland are largely unfilled. Their making was not followed by a dwindling flow of debris-laden meltwater incapable of keeping them flushed out. Flow seems to have ceased abruptly. For Grand Coulee's basins, this may have been caused by an abrupt failure of the Okanogan lobe dam but not for the hundreds of other empty basins in channels of all ages in the complex. If the scabland streams were supplied by lake water, instead of meltwater directly from the ice, they would have had only a minimum load. If they were supplied by a catastrophic failure of the Lake Missoula ice dam, an abrupt cessation would be expectable.

Ordinary valley trains along main scabland channels may have been deposited before flooding or in intervals between some floods. But, if so, these were largely destroyed by succeeding floods. None was made after the last flood to traverse any one channel.

All features of the Columbia plateau scabland are too intricately interrelated to dream of unlike agents or markedly different sequences for different parts. If any theory is correct for the Cheney-Palouse tract, it is correct also for Crab Creek, Grand Coulee, Quincy basin, Othello Channels, Pasco Basin, and the Columbia Valley farther downstream.

However plausible the fill-and-cut or ice-jam theories may be in any particular place, they must be equally applicable everywhere in the complex. The writers are confident that only the flood theory can successfully meet this test.

Recapitulation of Inferred History

(1) Miocene basalt flows (Columbia River formation) spread widely over, and almost completely buried, a mildly rugged topography of ancient crystalline rocks. Along the eastern and northern margins, bounded by highlands of these rocks, isolated peaks and ridges projected through the basalt and tongues of basalt penetrated back into deep valleys in the highlands.

(2) Lacustrine and fluvial sediments (Ellensburg, Latah, Early Ringold) were spread over at least parts of the region.

(3) Snake and Columbia rivers established courses across the basalt plain.

(4) Uplift with deformation down-tilted the region southwestward, warped and locally folded and faulted the basalt. These structures are strongly expressed in present topography. Snake and Columbia rivers held antecedent courses across some of the local uplifts.

(5) Weathering and erosion occurred.

(6) Later (?) Ringold sediment (Pleistocene lacustrine) was deposited on the eroded surface in the largest of the warped basins (Pasco), in southwestern part of the region.

(7) Loess was deposited on the plateau concomitant with continued erosional development of Columbia, Spokane and Snake valleys. Weathered zones and caliche layers in the loess record pauses during accumulation. Caliche developed also on basalt and sedimentaries where not loess-covered.

(8) Glaciation occurred. Canadian Cordilleran ice reached the northern edge of the plateau perhaps two or three times during the Pleistocene.

(9) Glacial water crossed the plateau when the deep Columbia valley along the high northern edge was blocked by lobes from northern highlands. Perhaps valley-train gravels were deposited in valleys across the plateau, their remnants now lying at or above upper limits of scabland gravels, or buried beneath them.

(10) Damming of Clark Fork and tributary valleys in Montana by a Cordilleran lobe in Purcell trench formed Glacial Lake Missoula. The Columbia valley was dammed by a similar lobe in Okanogan trench, presumably during Wisconsin time.

(11) The Lake Missoula dam failed before water had risen to the lowest col in Bitterroot Range. The lake's highest surface was about 4200 feet A.T., its depth close to the dam was 2000 feet. A recorded catastrophic emptying of Clark Fork portion demands similar emptying of the entire water body down nearly to 2500 feet A.T. This water escaped southwestward down Spokane Valley toward heads of scabland channels. Upper limits of channel heads range about 50 feet on either side of 2500 feet A.T.

(12) The great volume of glacial-lake water discharging down the dip slope of the plateau eroded its pre-existing minor valley system,

chiefly in loess, into a network of scabland buttes and rock basins in subjacent basalt. In deeper or broader places, deposits of basaltic and caliche gravel grew into huge bars. Some *de novo* divide crossings made by the flood became cataracts hundreds of feet high; Grand Coulee is the outstanding example. Depth and velocity (because of volume) enabled these rivers to quarry large columnar masses of essentially fresh basalt and transport them for miles. Giant current ripples left on bar summits indicate depth of rivers to have been considerably greater than heights of bars.

(13) Repeated reconstruction of dam with repetition of dam failures and consequent floods took place.

(14) Deepening of some channels in basalt during later floods allowed scarping of bars already formed and growth of lower bars close to scarp bases. These also have great current ripples. In Quincy basin, adjacent Columbia valley, and environs, seven successive floods are recorded, five of them while the Okanogan lobe dam forced the discharge to cross the plateau by way of scabland channels.

References Cited

Alden, W. C., 1953, Physiography and glacial geology of Western Montana and adjacent areas: U. S. Geol. Survey Prof. Paper 231, p. 1–193

Allison, Ira S., 1932, Spokane flood south of Portland, Oregon (Abstract): Geol. Soc. America Bull., v. 43, p. 133–134

—— 1933, New version of the Spokane flood: Geol. Soc. America Bull., v. 44, p. 675–722

—— 1941, Flint's fill-hypothesis for channeled scabland: Jour. Geology, v. 49, p. 54–73

Anderson, A. L., 1927, Some Miocene and Pliocene changes in northern Idaho: Idaho Bur. Mines and Geology, Pamph. 18

Bretz, J H., 1923a, Glacial drainage on the Columbia Plateau: Geol. Soc. America Bull., v. 34, p. 573–608

—— 1923b, Channeled scablands of the Columbia Plateau: Jour. Geology, v. 31, p. 617–649

—— 1925, The Spokane flood beyond the channeled scablands: Jour. Geology, v. 33, p. 97–115, 236–259

—— 1927a, Channeled scabland and the Spokane flood: Wash. Acad. Sci. Jour., v. 17, p. 200–211

—— 1927b, The Spokane flood; a reply: Jour. Geology, v. 35, p. 461–468

—— 1928a, Bars of channeled scabland: Geol. Soc. America Bull., v. 39, p. 643–702

—— 1928b, Channeled scabland of Eastern Washington: Geog. Rev., v. 18, p. 446–477

—— 1928c, Alternative hypotheses for channeled scabland: Jour. Geology, v. 36, p. 193–223, 312–341

—— 1929, Valley deposits east of the channeled scabland: Jour. Geology, v. 37, p. 393–427, 505–541

—— 1930a, Lake Missoula and the Spokane flood (Abstract): Geol. Soc. America Bull., v. 41, p. 92

—— 1930b, Valley deposits west of the channeled scabland: Jour. Geology, v. 38, p. 385–422

—— 1932, The Grand Coulee: American Geog. Soc. Spec. Publ., no. 15

Bryan, Kirk, 1927, The Palouse soil problem: U. S. Geol. Survey Bull. 790-B

Bucher, W. H., 1919, On ripples and related sedimentary surface forms: Am. Jour. Sci., v. 47, p. 149–210, 241–269.

Campbell, M. R., 1915, Guidebook of the Western U. S., Part A, The Northern Pacific Route: U. S. Geol. Survey Bull. 611, p. 142

Culver, H. E., 1937, Extensions of the Ringold formation: Northwest Sci., v. 11, no. 1., p. 57–60

Davis, W. M., 1921, Features of glacial origin in Montana and Idaho: Assoc. Am. Geog. Annals, v. 10, p. 75–147

Eakin, G. R., and Honkala, F. S., 1952, Cenozoic history of Missoula Valley, Missoula, Co., Montana (Abstract): Geol. Soc. America Bull., v. 63, pt. 2, p. 1361

Eardley, A. J., 1938, Yukon channel shifting: Geol. Soc. America Bull., v. 49, p. 343–358

Flint, R. F., 1936, Stratified drift and deglaciation of Eastern Washington: Geol. Soc. America Bull., v. 47, p. 1849–1884

—— 1937, Pleistocene drift border in Eastern Washington: Geol. Soc. America Bull., v. 48, p. 461–524

—— 1938, Origin of the Cheney-Palouse scabland tract: Geol. Soc. America Bull., v. 49, p. 461–524

Harding, H. T., 1929, Possible supply of water for the channeled scablands: Sci., v. 69, p. 188–190

Hobbs, W. H., 1943, Discovery in Eastern Washington of a new lobe of the Pleistocene continental glacier: Sci., v. 98, p. 227–230

—— 1947, The glacial history of the Scabland and Okanogan lobes, Cordilleran continental glacier: Privately printed, J. W. Edwards, Ann Arbor, 36 p.

Hodge, E. T., 1934, Origin of the Washington scabland: Northwest Sci., v. 8, p. 4–11

Jenkins, O. P., 1925, The Spokane flood; A discussion: Jour. Geology, v. 33, p. 747–748

Kerr, F. A., 1934, The ice dam and floods of the Talsekwe, British Columbia: Geog. Rev., v. 24, p. 643–645

Klebelsberg, R. V., 1948, Handbuch der Gletscherkunde und Glazialgeologie: Springer Verlag, Wien, v. 1, p. 212–218

Lane and Eiden, 1940, Sand waves in the lower Mississippi River: Western Soc. Engineers, v. 45, p. 281–291

Leighton, M. M., 1919, Road-building sand and gravel of Washington: Wash. Geol. Survey, Bull. 22, p. 200

Lupher, R. L., 1944, Clastic dikes of the Columbia Basin region, Washington and Idaho: Geol. Soc. America Bull., v. 55, p. 1431–1462

Matthes, G. H., 1948, Macroturbulence in natural stream flow: Am. Geophys. Union Trans., v. 28, p. 255–262

McKnight, E. T., 1927, The Spokane flood: a discussion: Jour. Geology, v. 35, p. 453–460
Nobles, L. H., 1952, Glacial sequence in the Mission Valley, Western Mont. (Abstract): Geol. Soc. America Bull., v. 63, pt. 2, p. 1286
Page, B. M., 1939, Multiple Alpine glaciation in the Leavenworth area, Wash.: Jour. Geology, v. 47, p. 785–815
Pardee, J. T., 1910, The glacial Lake Missoula, Mont.: Jour. Geology, v. 18, p. 376–386
— 1922, Glaciation in the Cordilleran region: Sci., v. 56, p. 686–687
— 1942, Unusual currents in glacial Lake Missoula, Mont.: Geol. Soc. America Bull., v. 53, p. 1569–1600
Piper, A. M., 1932, Geology and ground-water resources of the Dalles region, Ore.: U. S. Geol. Survey W. S. Paper 659, p. 130
Raisz, Erwin, 1941, Land forms of the Northwestern States (Map): Cambridge, Mass., The Author
Schwennesen, A. T., and Meinzer, O. E., 1918, Ground water in Quincy Valley, Wash.: U. S. Geol. Survey Water-Supply Paper 425, p. 131–161
Thiel, G. A., 1932, Giant current ripples in fluvial sediments: Jour. Geology, v. 40, p. 452–458
Trimble, D. E., 1950, Joint controlled channeling in the Columbia River basalt: Northwest Sci., v. 24, p. 84–88
Waters, A. C., 1933, Terraces and coulees along Columbia River near Lake Chelan, Wash.: Geol. Soc. America Bull., v. 44, p. 763–820

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SEISMIC EXPLORATIONS ON THE FLOOR OF YOSEMITE VALLEY, CALIFORNIA

BY BENO GUTENBERG, JOHN P. BUWALDA,* AND ROBERT P. SHARP

ABSTRACT

The depth and configuration of the bedrock floor beneath Yosemite Valley were determined by seismic surveys in 1935 and 1937. Seismic velocities of roughly 1.7, 2.5, 3.0, and 5.2 km/sec. and good to excellent reflections delineate at least three distinct layers within the valley fill resting on granitic bedrock.

The upper layer with a maximum thickness of about 150 m extends from Mirror Lake to the Wisconsin end moraines near Bridalveil Meadow. It is thought to be primarily deltaic lake deposits of Wisconsin age. The intermediate and basal layers have maximum thicknesses of 220 and 300 m respectively, and the intermediate layer lies in a U-shaped trough seemingly gouged out of the basal layer. Both layers are thought to be remnants of earlier lake fillings, and at least the basal layer is pre-Wisconsin. The greatest thickness of fill, about 600 m, is near the head of the valley between Ahwahnee Hotel and Camp Curry.

The bedrock floor of Yosemite Valley is an undulating surface with three separate basins and a total bedrock closure of at least 400 m, possibly approaching 500 m. The bedrock floor slopes steeply from the head of the valley to its deepest point, 600 m above sea level, between Ahwahnee Hotel and Camp Curry. Down-valley it rises rapidly about 300 m across a broad sill opposite Rocky Point. The second basin, 800 m above sea level, is opposite Cathedral Spires. From here the floor rises gradually down-valley to at least 1000 m above sea level opposite Artist Creek. It may rise another 100 m before the drop into a small basin more than 100 m deep at the Cascades. The amount of glacial excavation on the bedrock floor, essentially double the 450 m previously estimated, is attributed wholly to pre-Wisconsin glaciation. The greatest depth of excavation is in massive granitic rocks, and it is suggested that thick ice, exfoliation sheeting developed by pressure relief, and compressive flow in the glacier combined at this point to produce exceptionally effective erosion.

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* Deceased, August 19, 1954. This report has been prepared by Gutenberg except for the introduction and the section on geological interpretation written by Sharp. The views expressed therein do not necessarily represent the concepts of Buwalda which are known only from brief preliminary summaries (Gutenberg and Buwalda, 1938; Buwalda, 1941).

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Introduction and Summary

For fully 50 years earth scientists have differed strongly as to the efficacy of glacial erosion. One discussion of considerable warmth centered on Yosemite Valley, which Muir (1874a; 1874b; 1874c; 1874d) stoutly maintained was largely the product of glacier sculpturing and which Whitney (1865, p. 421–422; 1869, p. 81–87) contended was a product of tectonic subsidence. Other workers thought that erosion by running water and moving ice had done the job, but it remained for Matthes (1930, p. 31–32, 85–97) to show that these two agents played essentially equal parts in creating Yosemite Valley. Matthes' work is an outstanding attempt at a quantitative estimate of the magnitude of glacier excavation based on geological evidence.

The seismograph shows that even Matthes, an enthusiastic glacial sculpturist, underestimated by a full order of magnitude the amount of glacial excavation on the bedrock floor of Yosemite Valley. Where he estimated 450 m of deepening by glaciers, at least another 550 m must be added. Where Matthes (1930, p. 103) estimated the unconsolidated valley fill to be 90 m, it is closer to 600 m. Thus, the steep granitic walls rising 900 m above the present valley bottom are more than 1500 m above the bedrock floor. The visitor sees but three-fifths the splendor and magnitude of the bedrock relief in Yosemite Valley.

Yosemite glaciers did not everywhere cut to a uniform depth but left an undulating rock floor with broad depressions and rounded sills now deeply buried beneath unconsolidated Pleistocene fill (Fig. 8). The rock floor is 1100 m above sea level and buried beneath 100 m of fill near the head of Yosemite Valley. Downstream are three closed basins, the first and deepest being between Ahwahnee Hotel and Camp Curry. Its bottom is 600 m above sea level, and the fill is 600 m thick. The sill of this basin is 3.6 km down-valley, opposite Rocky Point, and has an elevation 900 m above sea level with an overburden 300 m thick. The second basin lying opposite Cathedral Spires has a bottom elevation of 800 m and a fill of 400 m. The slope is gently upward out of this basin at least as far as Artist Creek where the rock floor is at an elevation of 1000 m, and is buried by 160 m of fill. The crest of the sill between the second and third basins has not been accurately located. Incomplete evidence suggests that it may be as much as 100 m higher and possibly some distance down-valley from Artist Creek. The third basin lies opposite the Cascades, and the floor slopes up out of this basin at the exceptionally steep angle of 30°. The maximum closure on the bedrock floor of Yosemite Valley is at least 400 m, possibly closer to 500 m.

Yosemite Valley is thus an outstanding example of the efficacy of glacial excavation in overdeepening valleys and in creating large closed depressions. It is comparable in this respect to Lake Chelan, the Great Lakes, the Finger Lakes, and many fiords (Flint, 1947, p. 80–83, 90). The mechanism by which the glaciers were able to excavate closed basins of this magnitude remains a matter of speculation upon which some comments are offered later. In addition to valuable geological information, the seismological investigations in Yosemite Valley led to the recognition of basic relations significant to exploratory geophysics (Gutenberg, 1936; 1939). Field work on this project was begun in 1935 and continued into 1937 as a co-operative study by Gutenberg and Buwalda.

Acknowledgments

This study was made possible by the cooperation of the Yosemite Park Administration which arranged for permission to do the work, facilitated many aspects of the local logistics, and provided quarters for the field crews. J. L. Soske and R. A. Peterson participated in the field work and generously lent the equipment and services of their Geophysical Engineering Corporation. F. E. Lehner and A. B. Nomann served as operators of the seismic sets. Appreciation is also expressed for the assistance of the following men on the field parties: J. A. Ackerman, R. C. Doolittle, Martin Gould, J. F. Judson, K. G. Ku, S. P. Lee, E. F. Osborn, Sidney Shafer, and L. F. Uhrig. Eliot Blackwelder focused attention on the desirability of this type of work in Yosemite and visited the field operation. Financial support was provided by the California Institute of Technology and a grant in aid from the Penrose Bequest through The Geological Society of America.

Instruments

The seismic field outfit of the California Institute of Technology was designed by H. Benioff in 1932. It has been described by Gutenberg (1939). Four oil-damped seismometers of the variable-reluctance electromagnetic type having frequencies of approximately 40 cycles per second recorded on a Westinghouse Osiso Oscillograph with double string mirror galvanometers (natural frequency approximately 1500 cycles per second). Timing lines on the records are 0.01 second apart. Impulses from the firing circuit were usually recorded by the second instrument; they are marked by 0 on some of the seismograms which are reproduced. The first impulse, up, corresponds to the moment the current started to flow, the second impulse, down, to the interruption of the current at the moment of the explosion; the time, therefore, has to be counted from this downward impulse.

Parts of the outfit of The Geophysical Engineering Corporation which co-operated during the second period of the experiments have also been reproduced by Gutenberg (1939, Fig. 4, p. 175). Its 14 seismometers were of the variable-reluctance type with electromagnetic damping. The gain of the amplifier was automatically expanded at such a rate that the seismogram amplitudes remained within desirable limits. The galvanometers had a natural frequency high enough to insure uniform response up to 100 cycles per second. The output of several channels could be mixed in various ways. In routine field work, each channel recorded separately on one galvanometer first, and then other records were taken, utilizing one or two types of channel mixing.

General Data

At the beginning of the investigation it was expected that the depth to granitic bedrock in Yosemite Valley was nowhere greater than 300 m and that in extended regions it would be less than 100 m. With these conditions, the reflection method would not be of much use, since the reflected waves frequently would arrive too soon after the first motion to be recognizable on the seismograms. The refraction method was expected to yield all data desired. Consequently, a refraction profile was started southeast of the New Village, (Fig. 1, I). Surprisingly a high apparent velocity characteristic for granitic rock did not materialize with an increase of the distance between the instruments and the shot point until the instruments were south of the Ahwahnee Hotel, about 1.5 km from the shot point. When the same resulted by reversing the profile it was realized that the depth to granitic rock was greater than anticipated. Under these circumstances useful reflections could be expected, and reflections were recorded at the first shot points.

Instruments placed at a distance sideways from the long refraction profiles proved in several instances that certain pulses did not arrive from the direction of the shot point but from the granitic side walls of the valley. Since waves could also be reflected from the sides of the valley, frequently reflection records were taken with the line of the instruments in the direction of the valley as well as perpendicular to it. However, reflections from the side walls were rare. No reflections were obtained west of the Diversion Dam (Fig. 1, inset). Landmarks and the location of the main profiles are

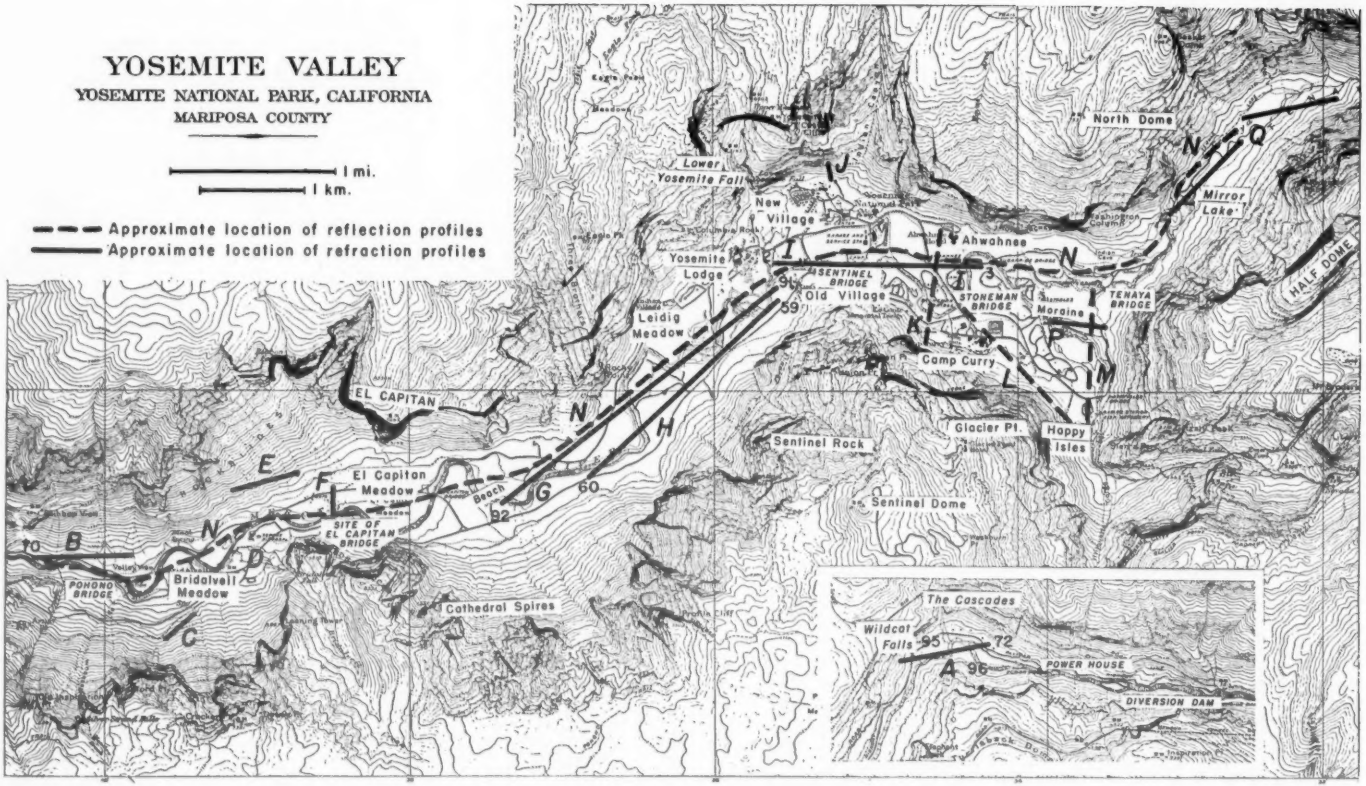


FIGURE 1.—MAP SHOWING LOCATION OF MAJOR REFLECTION AND REFRACTION PROFILES
Lettered from left to right

FIGURE 1.—MAP SHOWING LOCATION OF MAJOR REFLECTION AND REFRACTION PROFILES
Lettered from left to right

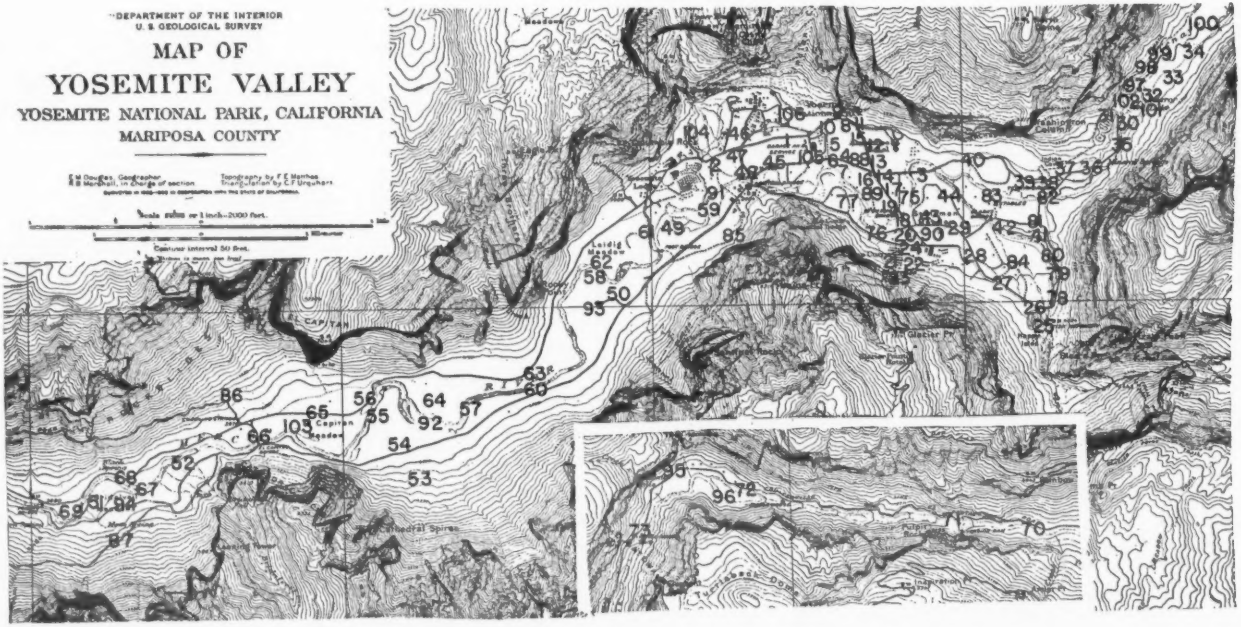


FIGURE 2.—MAP SHOWING LOCATION OF SHOT POINTS
Inset gives continuation of the main map to the west

TABLE 1.—DETAILS OF SHOT POINTS

Shot point no.	Direction and distance of instruments from the shot point In meters	Depth of charge Meters	Number of shots		Dynamite Pounds*
			Ground	Air	
1	W 32–816	1–3	7	6	38
2	E 120–1867	1–2	7	7	34
3	W 219–278 and 1266–2027; SE 219–278	1–2	8	4	33
4	W 43–141	2	3	1	Q
5	SE to SW 78–99	2	2	1	Q
6	E 43–145	2	1	1	Q
7	NE to NW 105–115	2	1	1	Q
8	W 29–115	2	4	1	Q
9	S 25–102	2	1	1	Q
10	N 44–118	2	2	1	Q
11	S 43–121	1	3	1	Q
12	N 58–103; S 135–198	1	7	2	1
13	N 155–200; S 59–119	1	10	2	1
14	N 273–317	1	1	1	Q
15	N 30 to S 30	1	2	1	Q
16	N 77–137; S 214–313; E 215–274	1	7	3	1
17	N 250–309; S 39–137	1	6	2	Q
18	N 199–297; S 182–276	1	5	2	1
19	N 95–191	1	2	1	H
20	S 59–153	1	2	1	Q
21	N 70–163	3	1	1	Q
22	N 143–239	1	4	1	1
23	N 41–115	1	6	1	Q
24	S 143–207	3	3	1	H
25	NW 156–236	1	2	1	H
26	NW 46–257	1	3	2	1
27	SE 108–216; NW 121–204	1	3	2	H
28	SE 140–221; NW 133–193	1	3	2	H
29	SE 122–211; NW 133–187; N 115–177	1	9	3	1
30	W 62–124	½	3	1	H
31	W 125–215	½	2	1	H
32	W 272–410	½	1	1	1
33	W 391–526	1	1	1	5
34	W 632–759	2	1	1	7
35	E 89–181	1	4	1	H
36	W 117–182	1	4	1	H
37	E 141–206; W 80–203	2	6	2	1
38	E 88–179; W 67–152	2	4	2	H
39	E 106–181; W 150–204	1	5	2	H
40	E 152–214; W 119–196	1	6	2	1
41	W 156–419	1	6	2	1
42	E 149–404	1	4	2	9
43	W 100–177; E 87–144	1	7	2	1
44	SE 101–164; NW 93–161	1	7	2	1
45	E 108–176; W 94–150	1	6	2	1
46	S 188–252	1	3	1	5
47	S 66–131	1	5	1	2
48	N 128–183	1	2	1	1
49	SE 100–153	1	3	1	H
50	N 92–161	1	4	1	2
51	ENE 109–192	1	4	1	H

TABLE 1.—*Continued*

Shot point no.	Direction and distance of instruments from the shot point In meters	Depth of charge Meters	Number of shots		Dynamite Pounds*
			Ground	Air	
52	E 176–224	1	3	1	1
53	N 59–117	1	3	1	Q
54	S 87–144; N 118–185	1	6	2	5
55	S 76–265	1	9	3	6
56	S 229–283	1	5	1	4
57	S 107–165	1	5	1	4
58	S 140–208; NW 106–149	1	10	2	7
59	W 157–2405	2	6	6	89
60	E 167–2222	2	6	6	113
61	N 98–150; S 84–144	1	4	2	H
62	N 72–124	1	4	1	H
63	W 125–183	1	2	1	H
64	N 71–122	1	4	1	H
65	E 77–133	1	4	1	H
66	N 92–141	1	4	1	H
67	N 107–130	1	1	1	Q
68	E 83–138	1	2	1	H
69	N 34–82	1	4	1	H
70	E 86–1075	½	7	3	29
71	E 112–160	½	3	1	1
72	W 116–854	1	7	4	34
73	W 46–126	1	3	1	2
74	E 105–880	1	2	2	10
75	W 65–114; E 99–217	1	12	4	2
76	E 66–134	11	4	2	2
77	NW 76–121; E 68–109	1	7	2	2
78	N 70–111	1	3	1	H
79	S 29–121; N 104–150	1	8	2	1
80	S 113–159; N 91–146	1	5	3	1
81	S 190–235; N 129–185	1	4	2	H
82	S 114–163; N 137–181	1	6	2	H
83	S 82–134	1	3	1	H
84	W 87–135	1	3	1	H
85	NE 80–120	1	4	1	H
86	W 63–679	1	2	3	11
87	SW 45–366	1	1	1	4
88	various directions up to 60 m	3	9	.....	1
89	SSE 153-NNW 153 + E 124–250	2	5	2	1
90	S 129 − N 137 + E 76–295	3	5	1	Q
91	NNE 140 − SSW 140; W 30 − 2800; S 50–200	3–16	11	3	81
92	ENE 156 − WSW 156; E 46–2845	3–10	9	4	145
93	WSW 137 − ENE 137 + N 120–278	12	6	1	Q
94	W 22–159 + N 106–288	3	8	1	1
95	E 21–826	4	6	2	51
96	SW 30–705; ENE 20–185	2	5	2	46
97	SW 62–457	3	2	1	1
98	SW 355–748	2	2	1	28
99	SW 509–897	2	1	1	31
100	SW 808–1211	3	1	1	52
101	SW 219–NE 186	1	2	1	1
102	SW 260–NE 147	3	2	1	9

TABLE 1.—*Continued*

Shot point no.	Direction and distance of instruments from the shot point In meters	Depth of charge Meters	Number of shots		Dynamite Pounds*
			Ground	Air	
103	W 122 − E 122 + NNW 63–225	9	8	1	Q
104	NNE 23–108 + SSE 18–110 + W 0–53	1	8	1	H
105	W 123 − E 122 + S 114–264; W 137–229; W 137 − NW 176 + N 132–231	3	15	3	2
106	N 76 − S 76 + WNW 92–206	1	6	1	H
Sum			480	180	948

* Q indicates one-quarter pound or less, and H between one-quarter and three-quarters pound. Other figures are given to the nearest pound.

indicated in Figure 1, the location of the shot points is shown in Figure 2; some numerical data for each shot point are given in Table 1.

The records at shot points 1–87 were taken between April 26 and May 6, 1935, by the instruments of the California Institute of Technology; those at shot points 88–106, between September 15 and 23, 1937, simultaneously by the instruments of the Geophysical Engineering Corporation and those of the California Institute of Technology. During the first period the ground was saturated with water almost to the surface; during the second the water table was at a depth of about a meter or more in most localities. During the first period vibrations of the ground caused by the water falls made it difficult to work close to them, especially near Yosemite Falls. In these areas the sensitivity of the instruments had to be reduced greatly. During the second period the water falls were almost dry, and recording conditions much better.

During refraction shooting at shot points 88–108 the 18 instruments of the two outfits were usually used in one line; during most reflection shooting at these points they were arranged along two perpendicular lines. In column 2 of Table 1 the first data for these shot points refer to the 14 instruments; the data following the sign + refer to the 4 instruments. Agreement between the two sets of instruments was good; frequently the reflections were somewhat clearer on the small outfit since this had characteristics which emphasized somewhat more waves with periods of less than 0.003 second than the large outfit. Waves reflected from shallow depths frequently have rather short periods.

At shot point 105 (near the garage and service station southeast of Park Headquarters) records were taken in various azimuths to provide a test on the direction from which the reflections arrived. In one instance the 14 instruments of the large outfit were arranged along two perpendicular lines the first 7 on a line extending from 137 m to 229 m west of the shot point, the other 7 on a line northward, also beginning 137 m west of the shot point.

During the experiments at the 106 shot points about 600 records of ground waves and about 200 of air waves (for determination of distance) were obtained. Twenty-five shot points were used mainly to get records of refracted waves, and at 74 shot points clear reflections were recorded. At the few shot points where no useful results of either kind were obtained the size of charges was limited either by nearness of houses or by the fact that large boulders made it impossible to secure with the available drilling facilities a shot hole even 1 meter deep. Holes for shot points 1–87, 97–102, and 104–106 were dug by hand, but for most holes at shot points 88–96 and at point 103 the drilling outfit of the Geophysical Research Corporation was used.

Distances at shot points 1–87 were determined by recording air waves from blasting caps or small charges; during the 1937 work (shot points 88–106) most distances of the instruments of the large outfit were measured with a tape, while most distances to the 4 instruments of the small outfit were calculated from the recorded sound waves. Comparison of results indicates that errors of 2 per cent were not unusual in calculations of distances from the recorded sound waves through the

air and that occasionally the error was 3 per cent.

Refraction Profiles

General Remarks

Travel times of refracted waves were determined on all seismograms and used for calculation of the velocity of longitudinal waves in the upper layer (not counting relatively thin surficial low-velocity layers). Results are plotted in Figure 4. East of the outermost end moraine (D, east of Bridalveil Meadow, Fig. 1) the velocity is about 1.7 km/sec., west of this moraine the velocity is about 2¼ km/sec. (Fig. 3). Shot points 51, 68, and 94 are close to the outermost moraine on the west, 52 is close to it on the east. The travel-time curve for shot point 66, along the moraine near the old El Capitan Bridge site, profile F (Fig. 4e) shows as low a velocity as the other shot points in that region.

Velocity in Moraines and Talus

To find data on the velocity in the moraines shot points 41 and 42 were selected with a reversed profile (Fig. 1, P) along the crest of the east moraine. The travel-time curve (Fig. 4e) indicates a layer about 40 m thick with a velocity of 1½ km/sec. on top of a layer with a velocity of 2.1 km/sec. Although up to 5 pounds of dynamite were exploded no wave with the high velocity characteristic of granitic rock was recorded up to a distance of 400 m from the shot point.

Travel-time curves (Fig. 4e) from shot points 86 in the talus north of the old El Capitan Bridge site (Fig. 1, E) and 87 in the talus slope south of Bridalveil Meadow on the old Wawona Road (Fig. 1, C) indicate a relatively well developed low-velocity layer on top. At shot point 86 this layer is underlain by material with a longitudinal velocity of 2.5 km/sec., while at 87 the corresponding velocity in the deeper material of the talus slope is 2.0 km/sec. Profile J extending on talus 76 m to the north from shot point 106 gave a constant velocity of 1.0 km/sec., indicating again that in the upper part of the talus slope the velocity is relatively small, probably owing to the porosity of this material near the surface and the absence of ground water.

Velocity in the Region of Mirror Lake

Six groups of refraction profiles were arranged along the Valley. For the groups in the region of Mirror Lake and northeast of it (Fig. 1, Q) shot points 30–34 and 97–102 were used. Since the recording trucks could not be moved to the east of the lake, shot points were selected there in such a way that travel-time curves in one direction are given by any shot point combined with all instruments, while for the reversed profiles any instrument may be used with all shot points. To find the order of magnitude by which the planes of the rays were tilted against the vertical plane, the instruments were about 100 m farther south at shot points 97–102 than at 31–34. Besides, some instruments were placed south and north of the profiles. The travel-time curves (Fig. 4a) found from the two groups of shot points are parallel except for small differences caused by conditions under the instruments. These differences were largest for the instruments on the lake bottom and those in the sand on the island in the eastern part of the lake, where the velocity in the surface layer is 1.1–1.2 km/sec. instead of 1.7–1.9 as it is west of the lake. The corresponding travel times have not been used in the calculations and are omitted from Figure 4a.

The travel-time curves for the Mirror Lake area indicate one or two layers with relatively small velocity and a deeper layer with higher velocity. The fact that the curves with the apparent high velocity are not strictly parallel is probably caused by small variations in the dip of the discontinuity. As the relatively large variations in the velocity of the upper layers do not permit high accuracy, average apparent velocities have been used in the calculations. The branch with the high velocity indicates the following apparent velocities (in km/sec.) southwestward (direction of dip):

Shot point.....	97	98	99	100	average
Apparent velocity..	4.7	3.7	3.9	4.4	4.2

In the opposite direction the apparent velocity is about 7 km/sec. If the relatively small effect of the low-velocity layer is neglected and a

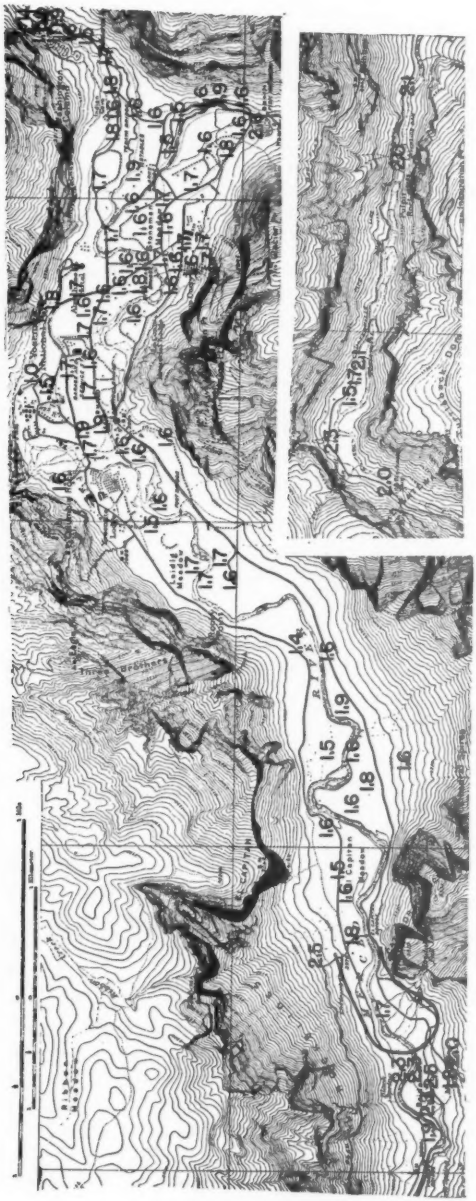


FIGURE 3.—VELOCITIES (IN KM/SEC) OF LONGITUDINAL WAVES NEAR THE SURFACE BELOW THE LOW-VELOCITY SUPERFICIAL LAYER

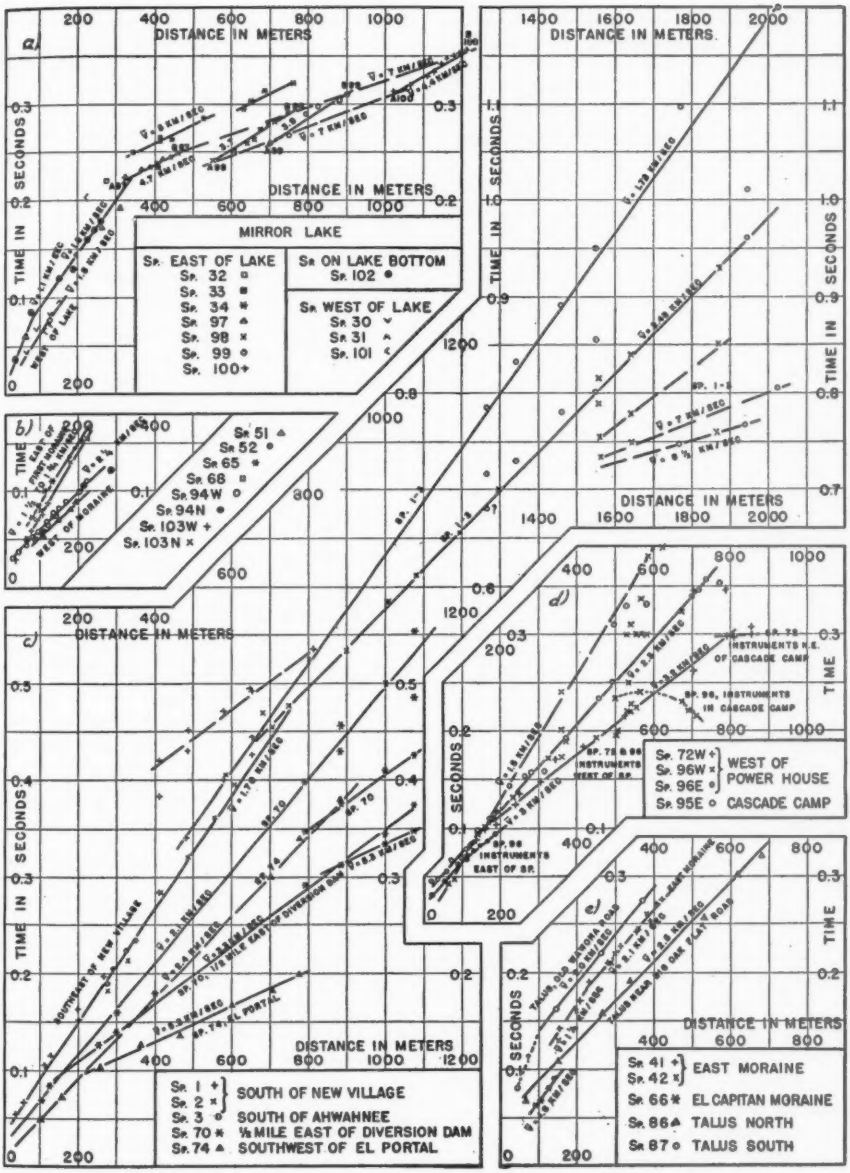


FIGURE 4.—TRAVEL-TIME CURVES ALONG REFRACTION PROFILES

velocity of 1.8 km/sec. is used for the surface layer, the average dip of the discontinuity is found to be about 5°. The velocity in the deeper layer is about 5.4 km/sec. This agrees well with a value of 5.25 km/sec. found earlier (Gutenberg, Wood, and Buwalda, 1932, p. 230) in granite[1] at the surface north of Yosemite Valley. Considering the slight differences in

[1] "Granite" is used throughout the paper in its widest sense and includes all granitic rocks.

dip indicated by the variations of the apparent velocities, the following depths to the granite are derived:

Shot point.....	102	97	98	99	100
Depth to granite....	130	110	80	60	35 meter

The dip seems to be less than 10° everywhere.

The good agreement between the results indicates that they give the correct order of magnitude. However, an error of 20 per cent seems possible, since the sand in the lake and east of it has a variable thickness which cannot be found without having instruments east of the lake, and since the difference between the velocity in this sand (1.1 km/sec.) and the normal surface layer (1¾ km/sec.) is relatively large. Moreover, the surface of the granite forms a channel dipping southwestward along the valley. Finally, the assumption of a layer with a constant velocity of about 1.8 km/sec. on top of the granite is a rough approximation. In most other regions of the valley this layer is separated from the granite by two or more layers in which the velocity of longitudinal waves is between 2 and 3½ km/sec. In the region of Mirror Lake the waves refracted through one or two such layers could arrive later than either the waves through the upper layer or those through the granite and could have escaped detection. If such layers with higher velocity exist, the depth to the granite would be slightly more than calculated.

Profile I Southeast of the New Village

Travel times along the refraction profile I between shot points 2 and 3 southeast of the New Village are about the same in both directions (Fig. 4c); therefore, the surfaces are roughly symmetrical to the center between the two shot points. Reflections indicate that the first (upper) discontinuity is approximately horizontal in the direction of the profile. The refraction profile indicates that its depth is about 150 m and that it separates a layer with a velocity of 1.7 km/sec. from a deeper layer with a velocity of 2.5 km/sec. At distances between 1.6 and 2 km from shot points 2 and 3, waves earlier than those through the second layer are recorded. Their high apparent velocities in both directions indicate that the corresponding layer in which their path reaches its deepest point is probably the granite. At shot point 3 it dips toward shot point 2, and at the latter it dips about the same amount toward shot point 3, so the floor seems to have a maximum depth between these two points. This may account for some of the difficulty in recording waves refracted through the granite along this profile. Moreover, there is the possibility that the refractions do not come from directly below. Finally, there may be a layer with a velocity of about 3 km/sec. on top of the granite, from which no refracted waves have been observed; their travel-time curve would probably intersect the curve corresponding to the velocity 2.5 km/sec. beyond the distance of 2 km. Consequently, the depth of the granite cannot be found reliably from this profile. It can be concluded that the top of the bedrock is at least several hundred m deep. Since charges of 20 pounds of dynamite have been used for records in both direction, it is not probable that earlier motion from a refracted wave has been missed.

Long Refraction Profiles G and H West of the Old Village

For the refraction profiles west of the Old Village, shot points 59 and 60 (profile H) have been used during the first period of the experiments and shot points 91 and 92 (G) during the second period. Since, in spite of the use of 30 and 43 pounds of dynamite, the seismograms recorded during the first period did not show waves with high velocity characteristic of granite at maximum distances of between 2 and 2½ km, this distance was increased to 2¾ km in the second attempt (shot points 91, 92), and charges of 50 and 60 pounds were used in both directions. Usually such charges are sufficient to record at distances over 5 km waves refracted at depths of several kilometers.

In some locations one instrument was used about 13 m south and another about 13 m north of the profile. At shot points 59 and 91 the first waves arrived from distances beyond 750 m with an apparent velocity of 4.4 km/sec. They recorded noticeably earlier on the southern instrument than on the instrument at the same distance on the profile, and still later

on the northern instrument. Thus, there is no doubt that these waves came from a direction with a considerable southern component, and not through the bottom of the valley under the profile.

In spite of relatively large amounts of dynamite, no travel-time curve indicating granite could be established for the profile extending to a distance Δ of 2800 m west of shot point 91. In the opposite direction (shot points 60 and 92) the first waves with a velocity higher than 1.6 km/sec. form a travel-time curve with an apparent velocity of 2½ and 3 km/sec. which intersects the travel-time curve of the direct waves at Δ = 750 m. Another curve through somewhat scattering points indicates an apparent velocity of 5 km/sec.; it intersects the curve of the direct waves at $\Delta = 1050 \pm$ m. A curve with shorter travel times gives an apparent velocity of 8 km/sec. or slightly more. It contains strong impulses at many distances and intersects the curve of the direct waves at Δ slightly less than 1000 m. The two most distant points of these two curves correspond to Δ = 2845 m with travel times of 0.859 sec. and 1.020 sec., respectively. Combining all these refraction data the only safe conclusions are that below the valley between Old Village and the Beach, north of Cathedral Spires, there is an upper layer with a velocity of about 1.6 km/sec. and a thickness exceeding 100 m and that the depth to the granite is in excess of 200 m.

Profile B Near Pohono Bridge

More definite results were furnished by the travel-time curve obtained at shot point 70, west of the Pohono Bridge with the instruments to the east. There is good indication for nearly horizontal layers along this profile which was not reversed. The travel-time curve (Fig. 4c) consists of three branches with apparent velocities corresponding respectively to 2.1 km/sec., 3.2 km/sec., and 5.2 km/sec. The resulting depth of the first discontinuity at shot point 70 is about 30 m, that of the bedrock somewhat more than 200 meters. Times of additional impulses form a travel-time curve with an apparent velocity of about 3.5 km/sec. or slightly more. If they indicate another layer, the depth of the granite would be slightly less but still close to 200 m.

Profiles Near A, South of the Cascades

The refraction profiles near A from shot points 72 and 96 west of the Power House (Fig. 2, inset) with the instruments toward the west and from shot point 95 south of the Cascades with the instruments toward the Power House indicate again a complicated situation (Fig. 4d). To the east of shot point 96 a layer with an apparent velocity of about 3 km/sec. extends almost to the surface, while to the west apparent velocities of 1.5, 2.3, and 3.2 km/sec. are observed. At shot point 95 a velocity of 2.3 km/sec. prevails immediately below the surficial low-velocity layer. An interesting feature of the travel-time curves is the fact that at certain distance intervals, between 600 and 900 m, the refracted longitudinal waves have an infinite apparent velocity (shot point 72) or arrive even earlier as the distance increases (point 96). In addition, these waves are recorded earlier at the instruments at shot point 96 than at corresponding distances from shot point 72 where the instruments were about 100 m farther north than for shot point 96. The best interpretation of the data from these shot points seems to be that south of the Cascades, where the valley changes direction (Fig. 2, left on inset) the granite slopes steeply eastward, and that there is also a moderate dip across the valley northward. The deepest point of the bedrock surface in this part of the valley is probably near the intersection of Cascade Creek and the road. The infinite apparent velocity indicates that the dip angle of the granite along the profile from 72 is about equal to the angle of incidence of the rays at the granite, which is roughly 30°. The maximum depth of the granite probably exceeds 100 m. Much more work would be required to determine more than the order of magnitude of the quantities involved in this complicated situation.

Profile South of El Portal Railroad Station

The last refraction profile (shot point 74, Fig. 4c) was about 10 km west of A in Figure 1, south of the El Portal railroad station. The granite, indicated by an apparent velocity of 5.2 km/sec., is at a depth of about 60 m; the velocity in the upper layer is 2.4 km/sec.

Results from the Refraction Profiles

The results obtained from the refraction profiles show that in the part of the valley east of the outermost end moraine (east of Bridalveil Meadow) a layer with a velocity of 1¾ km/sec. and a depth of the order of 100 m forms the upper part below the low-velocity surface layer. In most profiles the velocity in the deeper layer is about 2½ km/sec.; this is slightly more than that in the material near the surface west of the first moraine, and is about equal to the velocity in the talus. The velocity in the moraines is between 1½ and 2 km/sec. at the surface, but is slightly greater than 2 km/sec. at a depth of more than 40 m under the east moraine. The granite slopes from a depth of about 35 m at shot point 100 to about 130 m below Mirror Lake, and is at least a few hundred meters deep southeast of the New Village and east of El Capitan. Its depth is about 200 m west of Pohono Bridge. South of the Cascades the depth to the granite has a maximum in the northern part of the valley; near the turn of the valley the granite surface is inclined at an angle of roughly 30°. South of El Portal railroad station, about 10 km downstream from the Cascades, the granitic material is about 60 m below the surface.

Between the bedrock and the layer with a velocity of 2½ km/sec. there is, at least in some regions, a layer with a velocity of about 3 km/sec. Possibly still more discontinuities exist. Owing to the high velocity in the granitic bedrock either the waves through this or those through the upper layer are the first recorded in the refraction seismograms along the valley; waves through intermediate layers usually are obscured by the preceding motion.

In part of the valley between Ahwahnee Hotel and Bridalveil Meadow the refraction profiles do not furnish the accurate depth to the bedrock, since the waves refracted through the side walls or sloping portions on the sides make it difficult to interpret the seismograms. Besides, the length of the profiles required to record waves refracted through the granite as first impulse in the seismograms is so large that the effects of the topography of the granite surface are no longer negligible. Thus the interpretation of the travel-time curves becomes difficult without additional refraction profiles which would require many more data but would not guarantee better approximation.

Reflections and Depth of Discontinuities

General Remarks

In general, there was no serious difficulty in recording reflections where the distance of the instruments from the shot point was great enough to record the ground roll after the reflections. Since in Yosemite Valley the velocity of the ground roll rarely exceeds 300 m/sec., and usually reflections did not arrive after half a second, a distance of 100–150 m was usually sufficient to avoid obscuring the reflections by the ground roll.

Shot points 4–7 furnished the first reflection seismograms. For these shot points the instruments remained in the same position along a line approximately in the east-west direction and about 100 m long. Shot point 4 was east, 5 north, 6 west, and 7 south of the instruments. The agreement of the records proved that the reflections came practically from below. Along the lines 36–37–38–39–16 and 78–79–80–81–82 continuous profiling was used.

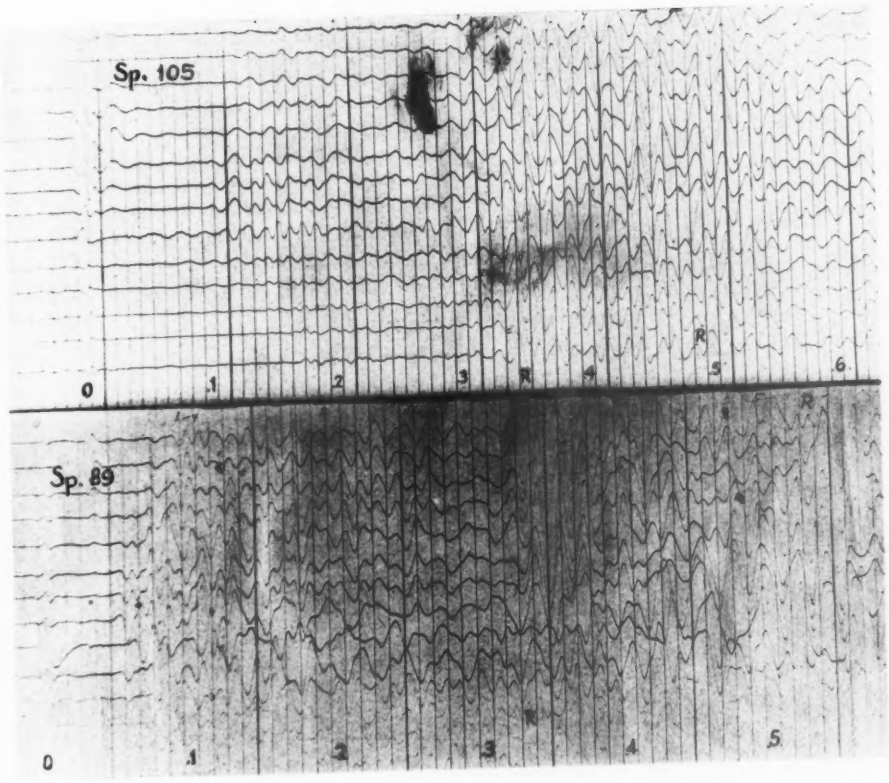
Most reflection seismograms show the same characteristics: the first waves are relatively small; near the shot point there is sometimes a small, occasionally a strong, reflection which has a travel time of about 0.1 second; after this the amplitudes decrease again. After a time interval of 0.1 or 0.2 second, depending on the region, the amplitudes increase suddenly in most seismograms, indicating a strong reflection. During the following 0.1 or 0.2 second the amplitudes remain large; occasionally other reflections are clearly indicated by outstanding waves arriving almost simultaneously at all instruments. This part of the seismogram ends

PLATE 1.—SEISMOGRAMS RECORDED AT SHOT POINTS 105 AND 89

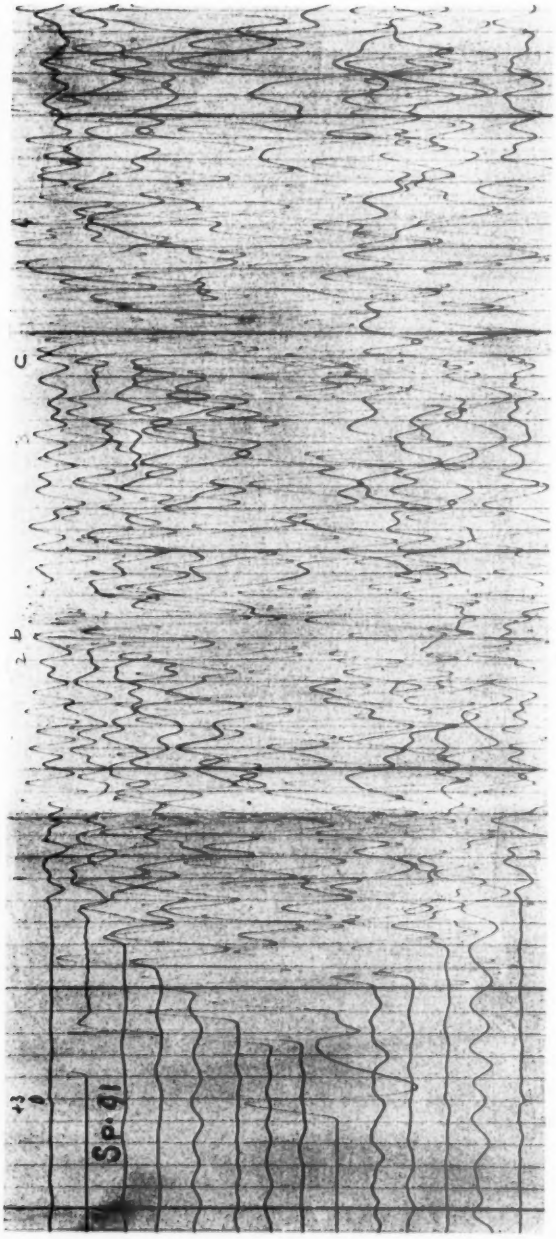
Instruments of Geophysical Engineering Corporation. At bottom: travel times in seconds. Light lines every 0.01 sec.

PLATE 2.—SEISMOGRAM FROM SHOT POINT 91

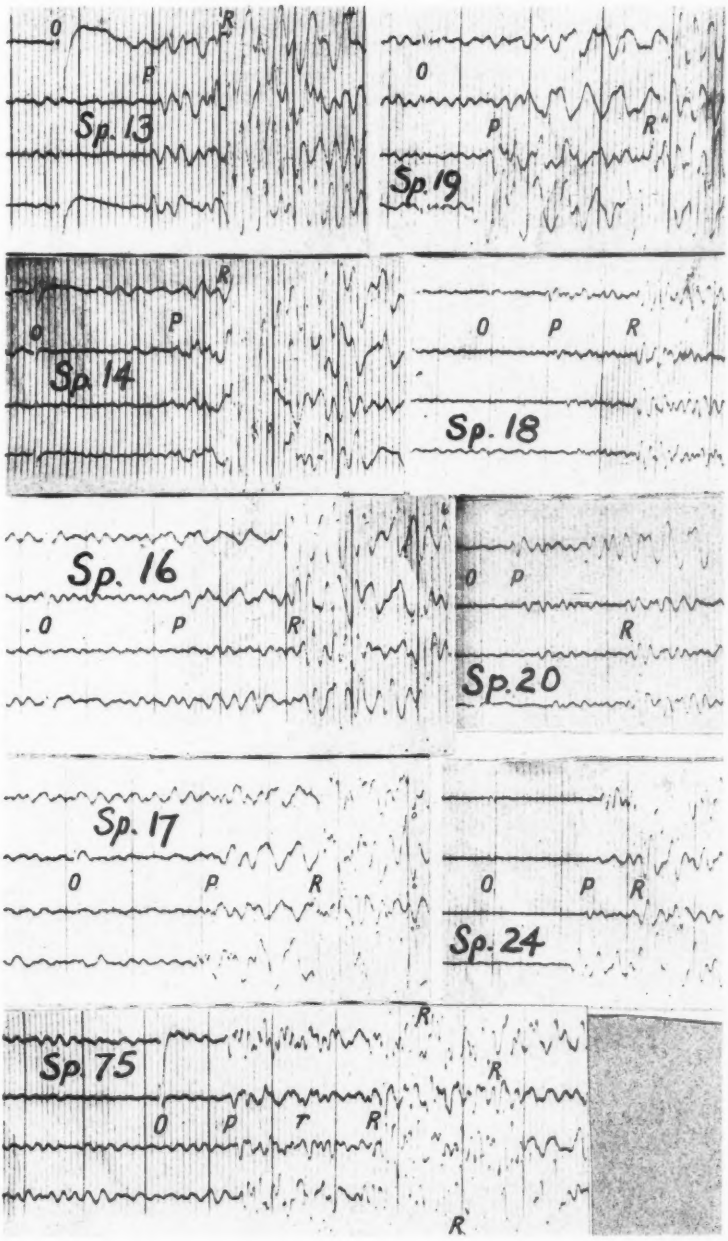
Details in Plate 1



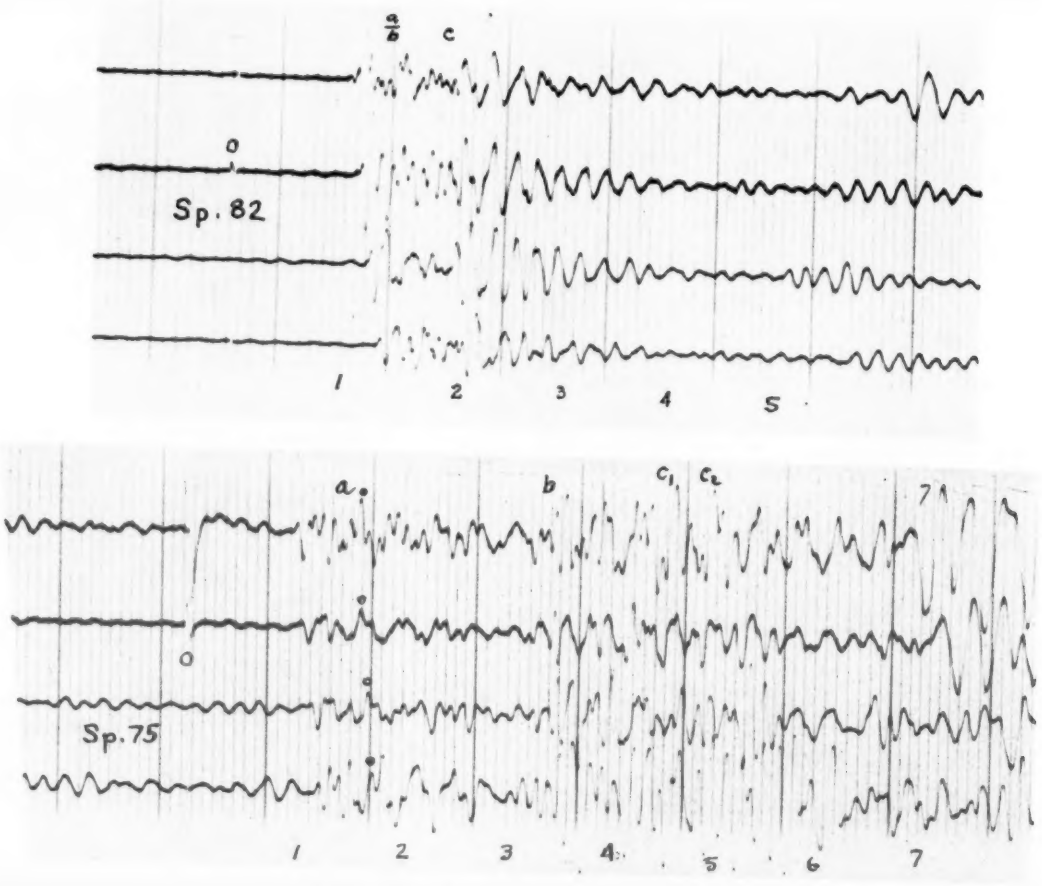
SEISMOGRAMS RECORDED AT SHOT POINTS 105 AND 89



SEISMOGRAMS FROM SHOT POINT 91



SEISMOGRAMS RECORDED ALONG NORTH-SOUTH PROFILE



SEISMOGRAMS FROM SHOT POINTS 82 AND 75

with one or more impulses, indicating a final reflection (*see* Plates 1, 2) after which the amplitudes decrease; waves with relatively small amplitudes and longer periods prevail until surface waves set in.

Correlation with Refraction Profiles

From comparison with results from the refraction profiles it follows that the first reflection comes from the discontinuity between the layers having velocities of about 1¾ km/sec. and about 2¼ km/sec., respectively. Since a slight increase of the velocity with depth is to be expected, average velocities of 1.8 and 2.5 km/sec. have been assumed for the calculations. The following deeper discontinuity produces the strongest reflections throughout the valley between the line Tenaya Bridge-Happy Isles to the east and at least to Pohono Bridge to the west. Near shot point 70 (Fig. 1) these reflections are probably produced by the discontinuity between the layers with velocities of 2.1 km/sec. above and about 3 km/sec. below. There may be layers with still higher velocities between this discontinuity and the surface of the granite.

Calculations of Depths and Dips of Discontinuities

From each seismogram the depth of the discontinuity at the point of reflection, its dip there (Table 2) and the horizontal distance of the point of the reflection from the shot point were calculated. No highly accurate results are to be expected, since the reflecting surface has an unknown shape; in addition the velocity in the various layers is known only approximately and its change with depth which controls the curvature of the ray is unknown. It was assumed that the rays are straight lines in each layer. The results are probably accurate within 20 per cent. Although no definite depth for the granite has been found from refraction shooting in the part of the valley where the granite is deepest, and though it has been assumed that the last clear reflection comes from the surface of the granitic layer, there is little doubt that the given depths to the granite in the central part of the valley are a good approximation, since to the east as well as to the west of these there is excellent agreement between the findings from reflections and those from refractions.

Column b of Table 2 contains the most and best reflections because of the usually sharp onset of larger movements at this point of the records (*see* Pls. 1, 2, 3, 4). Reflection *a* is frequently so early that it cannot be found in the movement produced by the direct and refracted waves; besides, its travel time changes fairly rapidly from one trace to the next so that its identification is sometimes doubtful. Reflection *c* is often obscured by the preceding movement and occasionally by the ground roll. In many records the reflection *c* may correspond to the rather sudden end of the short-period waves with relatively large amplitudes; the corresponding time, however, has not been used in this investigation, unless the waves on the traces lined up to indicate clearly a reflection. In some instances, the reflections *b* or *c* recorded poorly with small charges but were indicated by sudden onset of waves with large amplitudes and short periods (Pl. 2) if a larger charge was used. In connection with other seismograms, this frequently helped in identification of reflections; it is rarely now considered an indication of a reflection. Occasionally, reflections from a shallow depth record with greater amplitudes than the first direct or refracted waves (Gutenberg, 1936). The calculated dip for the layers *b* and *c* is given in Table 2; horizontal layering is indicated for the discontinuity *a* everywhere except at shot point 36 where a dip of about 25°W. is indicated by the records. No attempt has been made to find the direction or amount of the maximum dip.

Reflection Profiles

The two main reflection profiles, one (Fig. 1, K) across the valley from the west entrance

PLATE 3.—SEISMOGRAMS RECORDED ALONG NORTH-SOUTH PROFILE

By California Institute of Technology instruments. Light lines every 0.01 sec. O = time of explosion, P = first wave, R = reflection.

PLATE 4.—SEISMOGRAMS FROM SHOT POINTS 82 AND 75

a, b, and c are reflections, otherwise as in Plate 3

TABLE 2.—TIMES OF REFLECTION AND DEPTHS OF DISCONTINUITIES*

Shot point no.	Travel time of reflection in 0.01 sec.**			Depth of discontinuity at point of reflection			Dip†		Elevation of discontinuity††		
	(a)	(b)	(c)	(a)	(b)	(c)	(b)	(c)	(a)	(b)	(c)
1	....	28!	...	...	280	...	....	....	....	920	...
2	16?	26?	...	130	230	...	....	....	1070	970	...
3	26!	38!	38?	...	270	450	10W	....	....	950	770
4	....	30?	...	...	300	...	....	....	....	910	...
5	....	29!	...	...	290	...	....	....	....	910	...
6	....	30	...	...	300	...	....	....	....	900	...
7	14	30!	51?	120	300	600	....	....	1080	900	600
11	....	16?	...	...	150	...	11S	....	....	1060	...
12	11	21!	40?	90	190	400	12S	20S?	1110	980	750
13	14	27!	...	120	240	...	18S	....	1080	950	...
14	....	28!	42!	...	260	†††	15S	80N?	....	930	...
16	....	31!	43?	...	280	500	18S	18S?	....	900	660
17	....	34!	48?	...	310	550	6S	....	....	890	650
18	....	26!	38?	...	240	400	18N	....	....	920	700
19	11?	32!	...	90	290	...	17N	....	1110	900	(650)
20	....	24!	...	...	220	...	16N	....	....	960	(780)
21	....	17?	...	...	130	...	30N?	....	....	1050	(930)
22	....	16!	...	...	130	...	30N	....	....	1100	(1000)
23	....	....	...	...	...	...	....	....	....	(1140)	...
24	....	19!	27!	...	170	250	22N	35N?	....	990	820
26	....	....	[10?]	...	...	[90]	....	....	....	....	[1150]
27	....	10	...	...	90	...	0?	....	....	1130	...
28	13!	18!	26?	110	170	300	10NW	....	1110	1030	900
29	13!	25!	34!	110	250	380	0?	5SW?	1110	960	830
30	[11?]	....	18	[100]	...	160	....	....	[1140]	....	1090
31	....	....	18?	...	...	160	....	....	....	....	1090
35	....	[14?]	...	...	[120]	...	....	....	....	[1120]	(1080)
36	[10!]	[13]	17?	[70]	[120]	180	14W	....	[1160]	[1110]	1040
37	14	[14]?	21	120	[180]	220	5W?	5W?	1100	[1040]	1000
38	13?	....	22	120	...	220	5W?	5W?	1100	....	1000
39	13	....	20	120	...	200	5W?	5W?	1100	....	1020
40	13	[19?]	19?	120	[190]	...	5W	....	1100	[1030]	...
43	11?	29½!	40	90	300	450	0	0	1120	910	760
	14½!	....	...	120	...	...	17W	....	1190	....	...
44	9	28!	38	75	280	430	5NW?	....	1140	930	780
45	12?	22!	25	100	220	...	12E	....	1110	990	...
	....	25	...	...	250	...	....	....	....	950	...
47	13	26	37	115	260	440	....	....	1090	950	760
48	13	24!	38	120	240	450	3S	3S	1100	970	770
49	11?	[14!]	33?	90	[120]	450	....	....	1120	[1080]	870
	....	27	...	...	200	...	....	....	....	940	...
50	13	....	...	115	...	...	....	....	1100	....	...
51	....	12?	24!	...	110	260	?	23E	....	1090	910
	....	[20!]	...	...	[200]	...	20E	....	....	[960]	...
52	....	16!	26!	...	160	300	....	....	....	1030	890
55	8	17?	30?	70	180	330	0	....	1140	1020	870
56	....	19?	31?	...	180	340	0	....	....	1020	860
57	....	24	33?	...	240	360	10N	10N	....	970	850
58	....	22?	...	...	220	...	14N?	....	....	990	...
61	12!	[13]	30?	110	[130]	330	....	....	1100	[1080]	880

TABLE 2.—*Continued*

Shot point no.	Travel time of reflection in 0.01 sec.**			Depth of discontinuity at point of reflection			Dip†		Elevation of discontinuity††		
	(a)	(b)	(c)	(a)	(b)	(c)	(b)	(c)	(a)	(b)	(c)
62	11?	....	...	100	...	...	....	....	1010	....	...
63	....	....	30?	...	...	330?	....	....	....	....	880
64	14	[19!]	36	130	[200]	400	0	0	1080	[1010]	810
	....	[24!]	...	...	[250]	...	0	....	....	[960]	...
65	9!	23?	...	80	230	...	....	....	1120	980	...
66	....	15	25	...	150	290	9S	....	....	1050	910
68	....	10!	...	...	110	...	0	....	....	1080	...
69	....	5	16!	...	60	190	....	0	....	1120	990
70	....	....	14?	...	...	160	....	5W?	....	....	1010
75	10	32!	[47]	90	310	[550]	0	0	1110	890	650
	....	[42]	51	...	[460]	600	....	....	....	[740]	600
76	11	23!	...	100	230	...	0	....	1100	970	...
	....	31	...	...	330	...	....	....	....	870	...
77	11	24!	39?	90	230	450	10N	....	1110	970	750
	....	[30]	...	...	[300]	...	....	....	....	[900]	...
78	....	[5!]	...	...	[50]	...	....	....	....	[1190]	...
79	....	[7]	[10]	...	[70]	[100]	0	....	....	[1170]	[1140]
80	....	....	[11]	...	...	[110]	....	....	....	....	[1120]
81	12!	....	17	110	...	190	....	5S	1110	....	1040
82	12	....	18!	110	...	200	....	0	1100	....	1020
83	[9?]	[14?]	24?	[70]	[140]	280	0	....	[1160]	[1090]	950
84	9?	[11?]	[17?]	70	[100]	[200]	0	....	1170	[1130]	[1080]
85	8?	18!	...	65	130	...	0	....	1130	1070	...
88	....	25!	...	...	250	...	20S?	....	....	940	...
89	....	33!	52	...	320	600	5S	....	....	890	600
90	....	31!	38	...	310	430	0	....	....	900	780
91	11	20!	33!	100	200	400	17NE	6NE	1110	1010	810
92	16	28	35?	140	280	380	8W?	8W?	1060	930	830
93	12?	22?	28?	100	220	300	....	....	1120	990	910
94	....	....	26!	...	...	300	....	0	....	....	890
103	....	19?	29	...	190	330	....	5E	....	1010	880
104	8?	18	...	70	180	...	....	....	1130	1020	...
105	13	31!	45!	120	310	530	***	6E	1100	900	680
106	....	20?	25?	...	200	300	....	....	....	990	890

* Data in columns marked (a) refer to the discontinuity at the bottom of the layer with a velocity of about 1¾ km/sec.; in columns marked (b), to the discontinuity which corresponds to the beginning of the relatively large movements in the seismograms; in columns (c) to reflections which are considered to come from the granitic bedrock. Brackets indicate that the data may not refer to the discontinuity considered in the column. Some of these may belong to a reflection from a surface other than the discontinuities (a), (b), or (c).

** Travel times of reflections marked by "!" refer to strong, sharp impulses, those marked by "?" are not very clear, but there is little doubt that a reflection exists in that vicinity. Those not marked are at least good.

† Directions of dip refer to the component in the direction of the profile (Table 1, column 2). Question marks indicate that the numerical result is uncertain, but the order of magnitude and the direction of the dip are not questioned.

†† The elevation of a discontinuity at the shot point was interpolated if, owing to an appreciable dip, the point of reflection was far from vertical under the shot point. This is indicated by parentheses.

*** Dip of surface b at shot point 105 about 5°E. 100 m north of shot point dip about 11°S. 200 m south of shot point dip steeply northward.

††† Distance of reflecting surface 500 m; reflection probably from north side of valley.

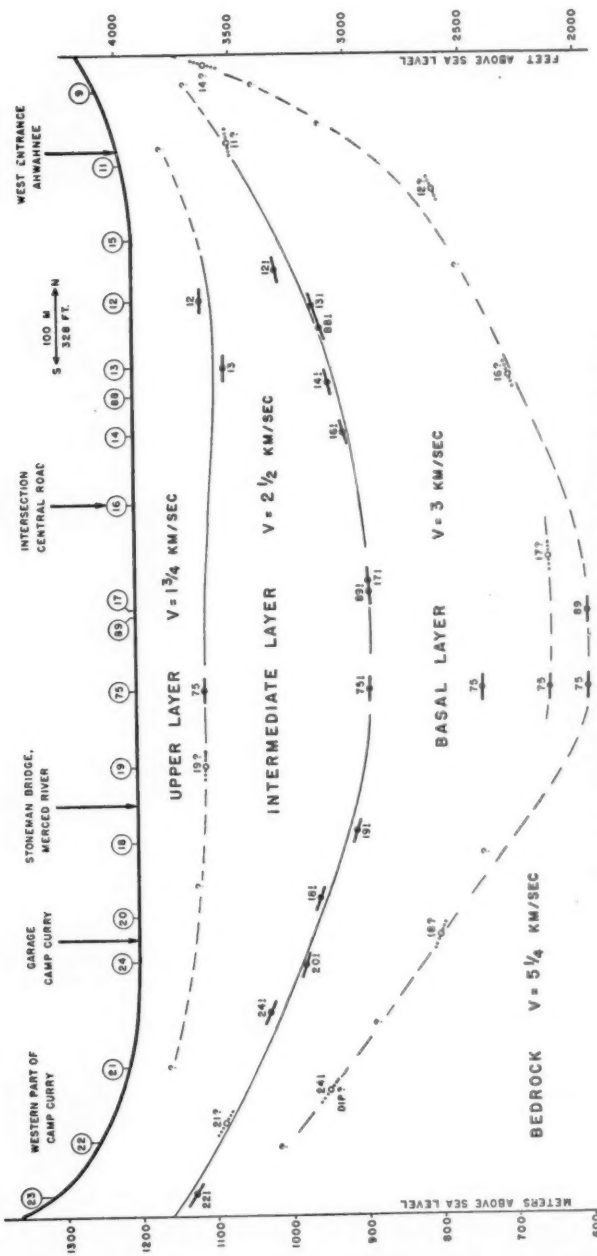


FIGURE 5.—PROFILE ACROSS YOSEMITE VALLEY NEAR CAMP CURRY
Numbers in circles indicate shot points

of Ahwahnee Hotel to the western part of Camp Curry and another (Fig. 1, N) along the valley from Mirror Lake to Pohono Bridge, are reproduced in Figure 5 and Plate 5, respectively. Other profiles include M (Fig. 1) along the eastern part of the valley from Happy Isles to Tenaya Bridge and L from Happy Isles to Stoneman Bridge.

Plate 3 shows several seismograms[2] recorded along the north-south profile K; the reflections *b* clearly show the increase in depth from shot point 13 (north) to 75 and the decrease from there to 24. Some show the reflection *c*. Seismograms reproduced on Plates 1 and 4, bottom (shot point 75), are examples of records from the deepest part of the valley. Large waves recorded at shot point 14 following the reflections marked R on Plate 3 indicate a practically horizontal arrival of the rays (dip of the reflecting surface 80°; Table 2, column c); probably they are reflected at the granite near the north wall.

Figure 4 was reproduced by Gutenberg (1939, p. 202) to show procedure of finding depth and dip of a discontinuity from reflections. A record from shot point 18 was reproduced by Gutenberg (1936, Fig. 3, p. 135) to illustrate dip calculations.

The main east-west profile is reproduced on Plate 5 in four sections. The eastern part of this profile was constructed from refraction data only. It is easily correlated with the following which is based on reflections. There is, however, some doubt about the proper correlation of the reflections near shot points 36 and 40/83. Where results at two shot points are plotted together on Plate 5, the point indicated by the upper figure is more to the north.

At shot point 105 on the second section AB, Plate 5, reflections were taken with the instruments in various directions. At the left end of this section the correlation of the discontinuity *b* is uncertain; the layer between *a* and *b* seems to thin out. It seems possible that the discontinuity *b* in the third section is not a continuation of *b* in the second. Otherwise, neither in this nor in the fourth section were correlation of reflections from point to point difficult, although distances between the shot points are greater than to the east. The outcrop of the discontinuity *a* between points 53 and 68 is taken from the observed velocities in the upper layer (Fig. 3). West of shot point 70 all information results from the refraction shooting.

Figure 6 shows contours of discontinuity *a* in meters, Figure 7 those of discontinuity *b*. However, west of Leidig Meadow (Fig. 1 near G) one or both materials separated by discontinuity *b* may be different from those separated by *b* east of Leidig Meadow. Figure 8 shows probable contours of the granitic bedrock (discontinuity *c*), which are based on all data discussed in this paper.

Surface Waves (Ground Roll)

On most records taken in Yosemite Valley surface waves were relatively large since conditions there were favorable for their development: soft, water-saturated layers near the surface and shallow depth of the explosives.

Areas with relatively small velocity of the onset of the ground roll stand out clearly (Fig. 9) and usually correlate with a relatively thin low-velocity weathering layer. Along the east wall of the valley (shot points 26, 78, 79, 80, 81, and 82) and to the east of Tenaya Bridge where the low-velocity layer was very thin or absent (Mirror Lake) the velocity of the ground roll is 300 or more m/sec.

The region with relatively high velocity of the ground roll in the lower part of the valley west of the first moraine agrees with the region where the velocity in the upper layer is relatively high. Frequently the velocity of the ground roll seems to be relatively high where there are large boulders. There are indications of two, and, in a few instances three, impulses of surface waves with different velocities.

Most of the observed waves of the gound roll have a period of about 0.03 sec. Periods between 0.05 and 0.08 sec. are rare, but longer waves with periods of almost 0.1 sec. have been recorded at several points. Most of the ground-roll waves have a length between 5 and 10 m; at 15 shot points waves with a length between 10 and 22 m occur, and at 3 shot points where

[2] For Plate 4 and similar figures seismograms obtained at a given shot point were selected which seemed to be best suited for reproduction; usually they were not those showing the clearest reflections. Many of these records obtained 20 years ago have faded.

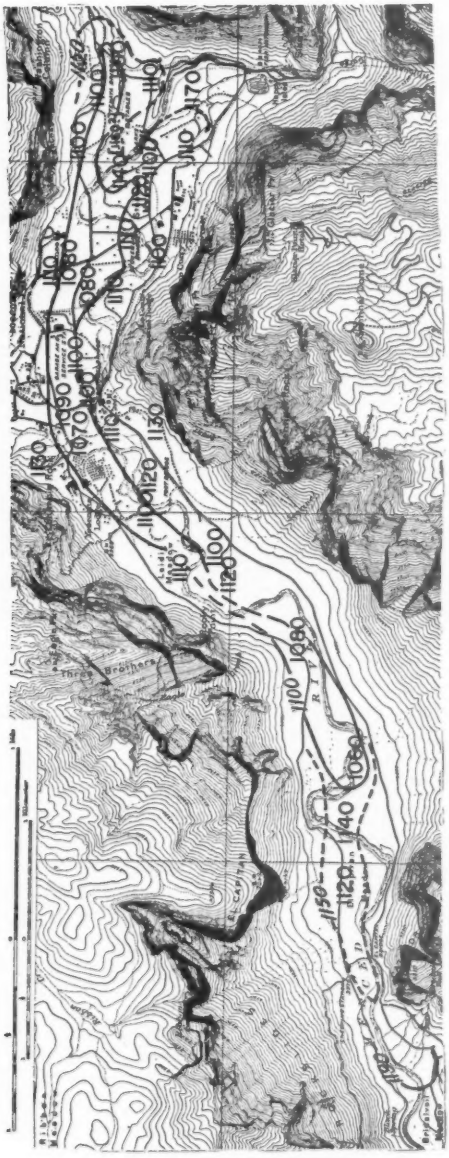


FIGURE 6.—CALCULATED ELEVATIONS (IN METERS) ABOVE SEA LEVEL OF DISCONTINUITY *a*
Elevation of the valley floor is about 1210 meters

FIGURE 7.—CALCULATED ELEVATIONS OF DISCONTINUITY *b*
Similar to Figure 6

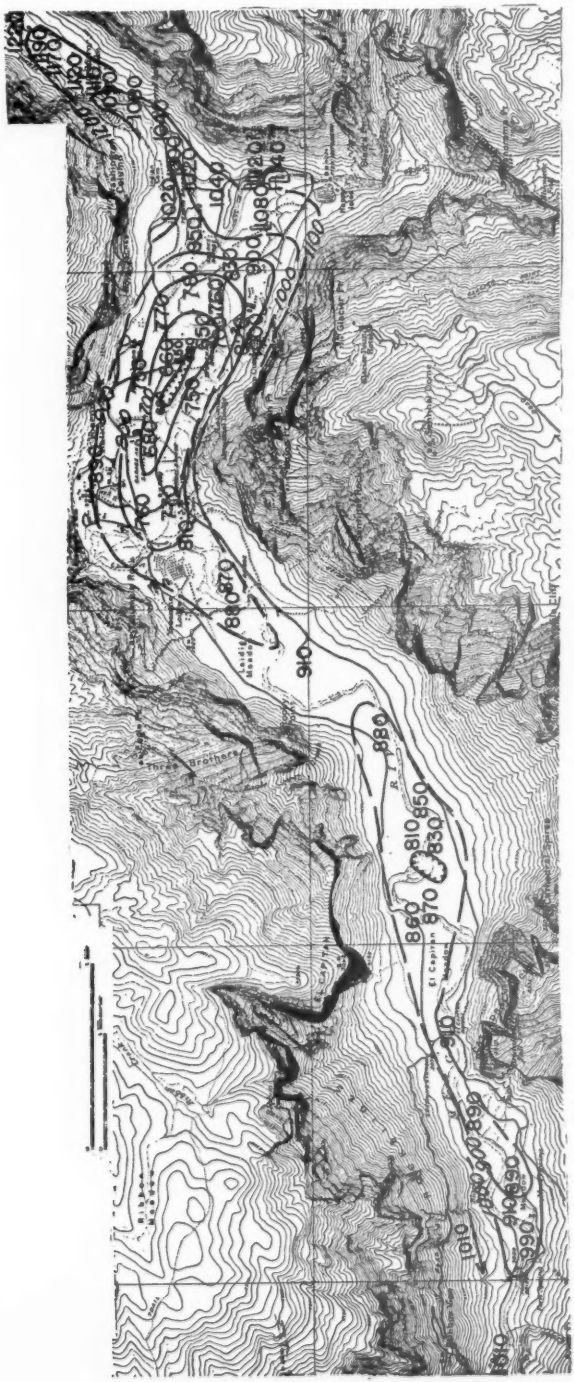


FIGURE 8.—CALCULATED ELEVATIONS OF DISCONTINUITY *c*
Supposed to be the bedrock. Similar to Figure 6

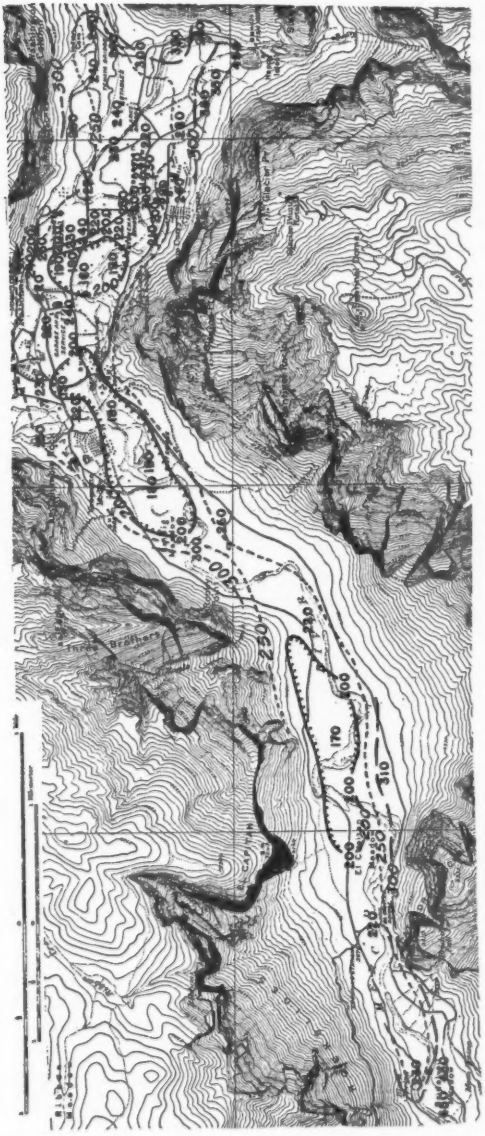


FIGURE 9.—VELOCITY OF GROUND ROLL IN METERS/SEC

long periods and a high velocity occur together the wave length was between 30 and 40 m.

Geological Interpretation

Erosional Features

Abrasion is not generally considered a major mechanism of glacial sculpturing in rocks of this type (Matthes, 1930, p. 89; Jahns, 1943, p. 94; Flint, 1947, p. 77; Holmes, 1949, p. 1430). Plucking or excavation of joint-bounded blocks is more favorably regarded, and most glacial geologists would probably join Matthes (1930, p. 90–94) and Zumberge (1955, p. 150) in placing strong emphasis on rock structure as a determining factor. This relation is supported by clear-cut field evidence at many places throughout the Yosemite region. However, the principle does not seem to apply in determining the location of the lowest point in Yosemite Valley between Camp Curry and Ahwahnee Hotel. There the valley walls are composed of some of the most massive least-jointed rocks in the area, and there is no reason to believe that the rocks underlying the valley floor are different. An intrusive contact between Sentinel granodiorite and Half Dome quartz monzonite crosses the valley at about this point (Calkins, 1930, Pl. 51), but it probably has little structural significance for it is not easily discerned on the valley walls and has not been etched out by weathering or erosion.

A more likely factor was merging of the Tenaya and Merced glaciers, the two principal ice streams composing Yosemite Glacier. The location of the deepest point somewhat west of the junction of these two canyons is probably related to the greater thickness of the ice there, at least 1700 m, and to the mode of flow within the glacier. At some stage in its evolution the floor of Yosemite Valley must have had a gentler gradient than the floors beneath Tenaya and Merced glaciers. This tended to cause a decrease in flow velocity of the ice which in turn lead to compressive flow as described by Nye (1952, p. 86–89). Compressive flow sets in wherever the flow velocity decreases, and it involves a surfaceward movement of the ice which can lead to an increase in thickness. This change was probably most marked a short distance below the junction, and it was here that ice thickness and surfaceward movement were the greatest. Surfaceward movement would seem to favor excavation by providing a mechanism for removing debris from the floor and for carrying it up and out of any depression created by erosion. As the difference in rock floor gradients became more marked the tendency for compressive flow increased and presumably excavation was favored.

This is but a partial answer, for it does not tell how the glacier was able to loosen blocks in such massive, sparsely jointed rocks. One good possibility is that exfoliation sheets developed through the release of pressure by erosion as suggested by Jahns (Personal communication) and described by Lewis (1954, p. 418–422). Exfoliation sheets are widely displayed in these granitic rocks, and their separation may have been promoted by pressure-controlled subglacial freeze and thaw (Holmes, 1944). Furthermore, the amount of glacial excavation was probably more than sufficient to cause sheet jointing even while the rocks were deeply buried by ice. The combination of sheet-joint development in rocks and compressive flow in glaciers should be particularly effective in excavation, as these are to some degree self-enforcing processes that gain in strength as they progress.

The shallow basin north of Cathedral Spires can be attributed to excavation of well-jointed rocks as described by Matthes (1930). The rocks on the south side of the valley and, at least in part, underlying it, consist of an interpenetrating complex of diorite, gabbro, El Capitan granite, Bridalveil granite, and Taft granite (Calkins, 1930, Pl. 51), all more or less well jointed. Rock structure appears a reasonable and adequate explanation for the location of this basin, as no tributary glaciers of any size joined the trunk glacier in this vicinity.

The sill separating the first two basins lies opposite Rocky Point which is an apophysis of massive El Capitan granite that extends across the valley. The sill is presumably held up by this relatively resistant sparsely jointed rock. The gradual rise of the valley floor downstream from the second basin is attributed to decreasing excavation related to a decreasing ice thickness (Buwalda, 1941, p. 92), for the rocks in the walls of that part of the valley are rela-

tively well jointed. The principal episode of bedrock excavation on the floor of Yosemite Valley must be of pre-Wisconsin age as previously recognized by Matthes (1930, p. 74), but just which of the pre-Wisconsin glacial stages were involved is not known.

Depositional Features

The nature and origin of the three principal layers within the 600 m Yosemite Valley fill delineated by the seismic explorations are of interest. These layers are directly related to the glacial history which includes at least three, probably more, separate episodes of glaciation. Matthes (1930, p. 73–75) recognized a Wisconsin stage, represented by a group of six moraines on the valley floor between Bridalveil Meadow and the old El Capitan Bridge site, and two pre-Wisconsin stages, El Portal (probably Illinoian and Glacier Point (either Kansan or Nebraskan). Blackwelder (1931, p. 907–909) feels that the moraines near Bridalveil Meadow are most likely early Wisconsin (Tahoe) and that the El Portal and Glacier Point stages are parts of a single glaciation correlative with the Kansan (Sherwin) stage on the east slopes of the Sierra Nevada Range. The moraines on the floor of Yosemite Valley appear to be older than late Wisconsin, and Blackwelder (1931, p. 907) is probably right in stating that late-Wisconsin (Tioga) glaciers did not reach the floor of the valley. J. H. Birman (Personal communication), on the basis of studies in the Yosemite region and of moraines on the west slope of the Sierra Nevada Range, tentatively considers the moraines near Bridalveil Meadow to be about middle Wisconsin, correlative with a similar stage on the South Fork of San Joaquin River (Birman, 1954, p. 42). He is inclined to regard the El Portal stage as early Wisconsin rather than pre-Wisconsin. It is clear that glaciers moved into the area at least three and in all probability four times, since four stages of glaciation are known on the east slope of the Sierra Nevada Range (Blackwelder, 1931, p. 918). However, Matthes' terminology of Glacier Point, El Portal, and Wisconsin will be followed.

The uppermost layer in the valley fill with a velocity of 1.7 km/sec. and a maximum thickness of about 150 m extends from the head of the valley to the outermost Wisconsin end moraine just east of Bridalveil Meadow. This layer is clearly related to and somewhat younger than the Wisconsin moraines. It is thought to be composed chiefly of gravel, granitic gruss, and sand deposited largely as deltaic beds in the lake impounded behind the moraines. This lake was at least 9.5 km long and 150 m deep. Some of the filling may be of late-Wisconsin age, representing debris washed out from late-Wisconsin (Tioga) glaciers that did not reach the floor of the valley.

One puzzling feature is the essential horizontality of the contact (discontinuity *a*) at the base of the upper layer as it appears in a cross-valley section (Fig. 5). One would expect the Wisconsin glacier to have scoured out a more distinct U-shaped trough in the underlying materials. Data on the attitude of this contact near the valley margins are lacking, so possibly the trough is gently U-shaped and the interpretation of the upper layer as a lake filling in a basin created by ice scour and dammed by the moraines is permissible. The 25-degree westward dip of discontinuity *a* at station 36 near the mouth of Tenaya Valley may mark the position of a buried Wisconsin moraine. Matthes (1930, p. 57–58) considers the medial-like moraine projecting from the wall at the head of Yosemite Valley to be part of a Wisconsin end moraine; possibly other smaller moraines are buried in the lake fill.

The intermediate layer in the valley fill with an average velocity of 2.5 km/sec. and a maximum thickness of about 220 m lies in a U-shaped trough (discontinuity *b*) presumably gouged out of the basal layer of the valley fill by the Yosemite Glacier. The intermediate layer extends beyond (west of) the Wisconsin moraines near Bridalveil Meadow where it becomes the surface layer (Pl. 5) and has a lower velocity, 2.1 km/sec. The suggestion has been made that both the intermediate and basal layers in the valley fill are till (Gutenberg and Buwalda, 1938) because the seismic velocities are about right, but it seems more likely that they are remnants of earlier glacial-lake fillings. Thick tills are uncommon on the floors of glaciated mountain valleys, even though overdeepened. Furthermore, glacial excavation into the basal layer created a closed basin at least

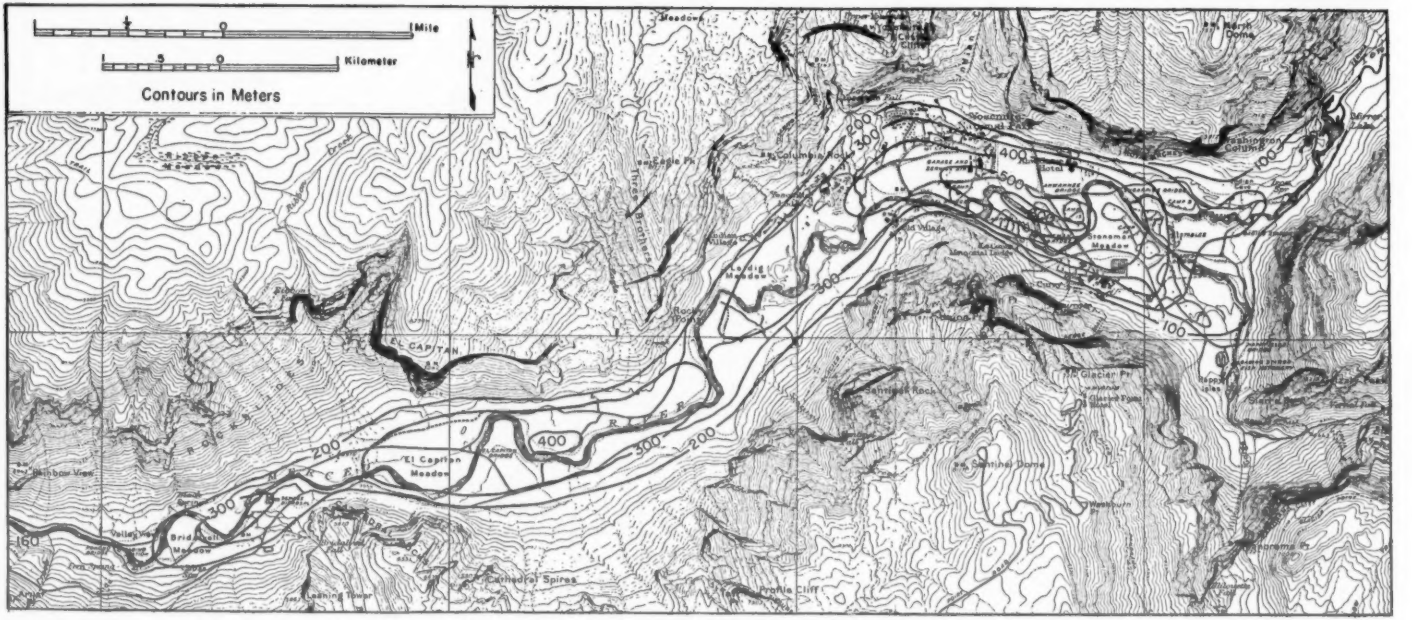


FIGURE 10.—DEPTH OF PLEISTOCENE FILL IN YOSEMITE VALLEY IN METERS

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250 m deep, and such a basin was more likely to be occupied initially by a lake than by a deposit of till. Therefore, it is postulated that the intermediate layer in the valley fill represents deposits, largely deltaic, laid down in a lake created by Yosemite Glacier. The fact that this layer extends beneath and beyond the Wisconsin moraines indicates that it is older and probably related to El Portal stage. Its seismic velocity is greater than that of the upper layer, and this is attributed in part to compaction through overriding by the Wisconsin glacier. This is confirmed by the lower velocity west of the Wisconsin moraines, although the compaction there owing to age is great enough to give a velocity (2.1 km/sec.) which is greater than that of the Wisconsin lake fill (1.7 km/sec.).

The basal layer in the valley fill with a velocity of 3 km/sec. and a maximum thickness approaching 300 m rests in most places on the granitic floor. It too is probably a lake filling formed after the glaciation which excavated the bedrock. This may have been the Glacier Point or a still older stage. These deposits were deeply scoured by ice of El Portal stage, and the greater seismic velocity is probably an expression of compaction related to greater age and to the weight of the relatively thick El Portal ice.

As shown in Plate 1 and as reported in the preceding text, layers or discontinuities other than those treated above may exist within the valley fill, but they are too ill-defined to warrant discussion. A thin deposit of till may lie along the contact between the intermediate and the basal layers and perhaps also at the base of the upper layer.

The difference in velocity of the upper layer west (1.7–1.9 km/sec.) and east (1.1–1.2 km/sec.) of Mirror Lake at the mouth of Tenaya Valley is an interesting anomaly. The floor of Tenaya Valley above Mirror Lake is filled with loose granitic gruss and sand which is probably in part a delta built into the lake and in part a channel fill laid down by the stream. The composition of this material may be the same as the valley fill west of the lake, but it is less compacted and of lower velocity owing to a much younger age. Matthes (1930, p. 105) shows that Mirror Lake, hence, at least part of the fill to the east, is a post-glacial feature created by large rock avalanches.

The relations recorded by refraction shooting between the Powerhouse and the Cascades in the western part of the valley are complex, and the data are too few to permit reliable interpretations. The depression centering south of the Cascades is too deep, 100 m or more, to be of any but glacial origin. The layers and lenses of debris filling this depression, velocities of 1.5, 2.3, and 3.2 km/sec. may be of several origins. Some are probably lake fill and some are probably glacial outwash and stream gravels. Judging from surface exposures the 60 m of fill in the valley bottom near El Portal Railroad station, 10 km down the canyon, is probably bouldery outwash and stream gravel.

References Cited

Birman, J. H., 1954, Pleistocene glaciation in the upper San Joaquin Basin, Sierra Nevada: Calif. Div. Mines Bull. 170, Chap. 5, p. 41–44

Blackwelder, Eliot, 1931, Pleistocene glaciation in the Sierra Nevada and Basin Ranges: Geol. Soc. America Bull., v. 42, p. 865–922

Buwalda, J. P., 1941, Form and depth of the bedrock trough of Yosemite Valley: Yosemite Nat. Notes, v. 20, p. 89–93

Calkins, F. C., 1930, The granitic rocks of the Yosemite region: U. S. Geol. Survey Prof. Paper 160, 137 p.

Flint, R. F., 1947, Glacial geology and the Pleistocene epoch: N. Y., John Wiley and Sons, 589 p.

Gutenberg, Beno, 1936, On some problems concerning the seismic field methods: Beitr. angew. Geophys., v. 6, p. 125–140

—— 1939, Zur Entwicklung der seismischen Aufschlussmethoden: Gerlands Beitr. Geophys., 4. Supplem., Ergebnisse d. kosm. Physik, v. 4, p. 169–218

Gutenberg, Beno, and Buwalda, J. P., 1938, Geophysical Investigation of Yosemite Valley (Abstract): Geol. Soc. America Proc., 1937, p. 240

Gutenberg, Beno, Wood, H. O., and Buwalda, J. P., 1932, Experiments testing seismograph methods for determining crustal structure: Seismol. Soc. America Bull., v. 22, p. 185–246

Holmes, C. D., 1944, Hypothesis of subglacial erosion: Jour. Geology, v. 52, p. 184–190

—— 1949, Glacial erosion and sedimentation: Geol. Soc. America Bull., v. 60, p. 1429–1436

Jahns, R. H., 1943, Sheet structure in granites: its origin and use as a measure of glacial erosion in New England: Jour. Geology, v. 51, p. 71–98

Lewis, W. V., 1954, Pressure release and glacial erosion: Jour. Glaciology, v. 2, p. 417–422

Matthes, F. E., 1930, Geologic history of the Yosemite Valley: U. S. Geol. Survey Prof. Paper 160, 137 p.

Muir, John, 1874a, Mountain sculpture: Overland Monthly, v. 12, p. 393–403

—— 1874b, Mountain sculpture, origin of Yosemite valleys: Overland Monthly, v. 12, p. 489–500
—— 1874c, Ancient glaciers and their pathways: Overland Monthly, v. 13, p. 67–79
—— 1874d, Glacial denudation: Overland Monthly, v. 13, p. 174–184
Nye, J. F., 1952, The mechanics of glacier flow; Jour. Glaciology, v. 2, p. 82–93
Whitney, J. D., 1865, Geological survey of California: Geology, v. 1, 498 p.
—— 1869, The Yosemite guidebook: Geol. Survey Calif., 155 p.
Zumberge, J. H., 1955, Glacial erosion in tilted rock layers: Jour. Geology, v. 63, p. 149–158

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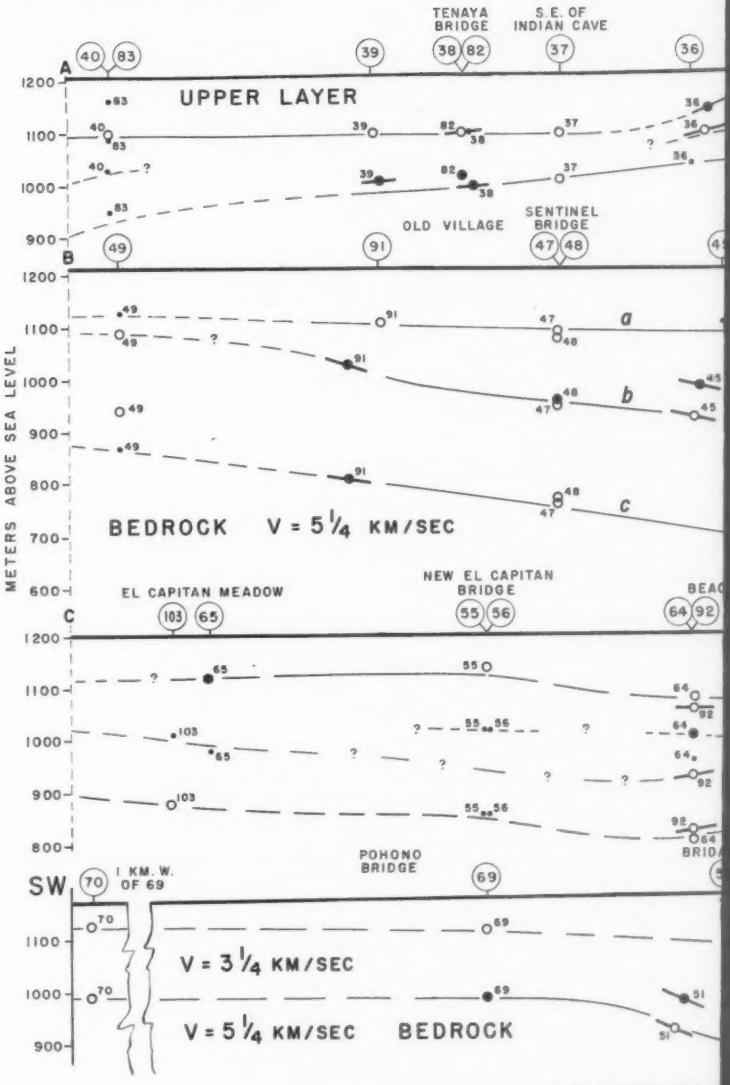
TENAYA BRIDGE
S.E. OF INDIAN CAVE
A
UPPER LAYER
OLD VILLAGE
SENTINEL BRIDGE
B
a
b
c
BEDROCK V = 5 1/4 KM/SEC
METERS ABOVE SEA LEVEL
EL CAPITAN MEADOW
NEW EL CAPITAN BRIDGE
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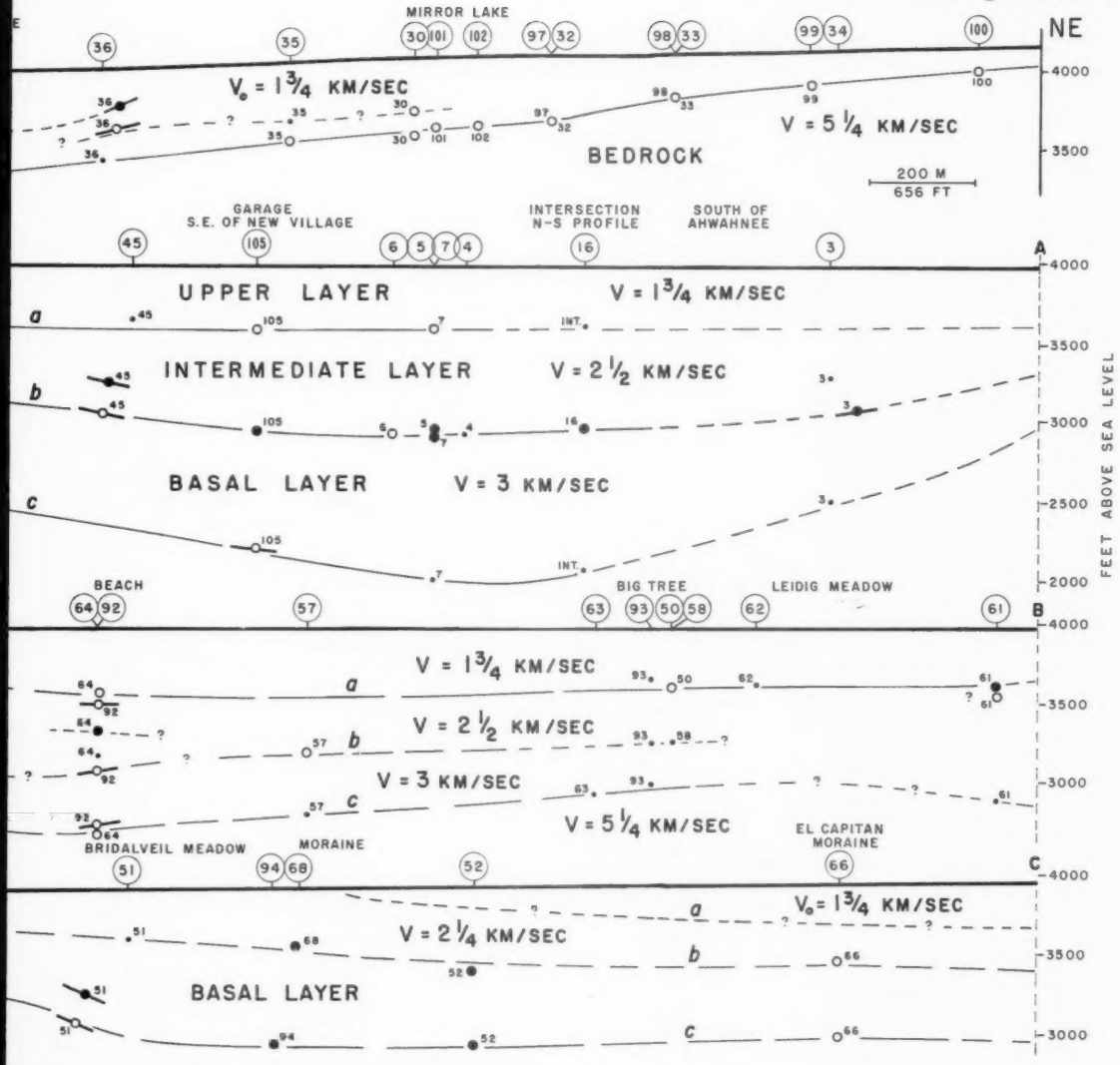
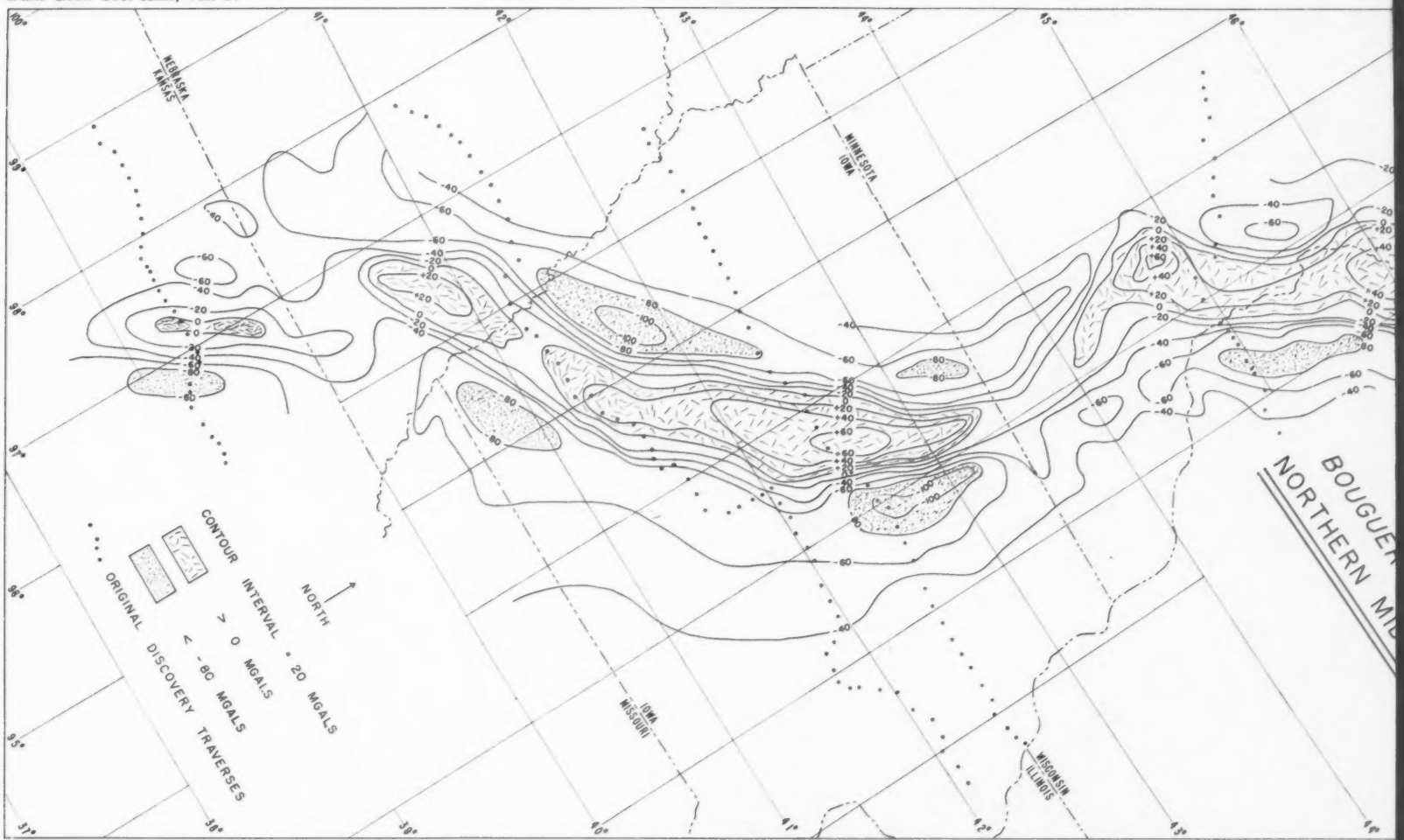
1 KM. W. OF 69
POHONO BRIDGE
V = 3 1/4 KM/SEC
V = 5 1/4 KM/SEC BEDROCK

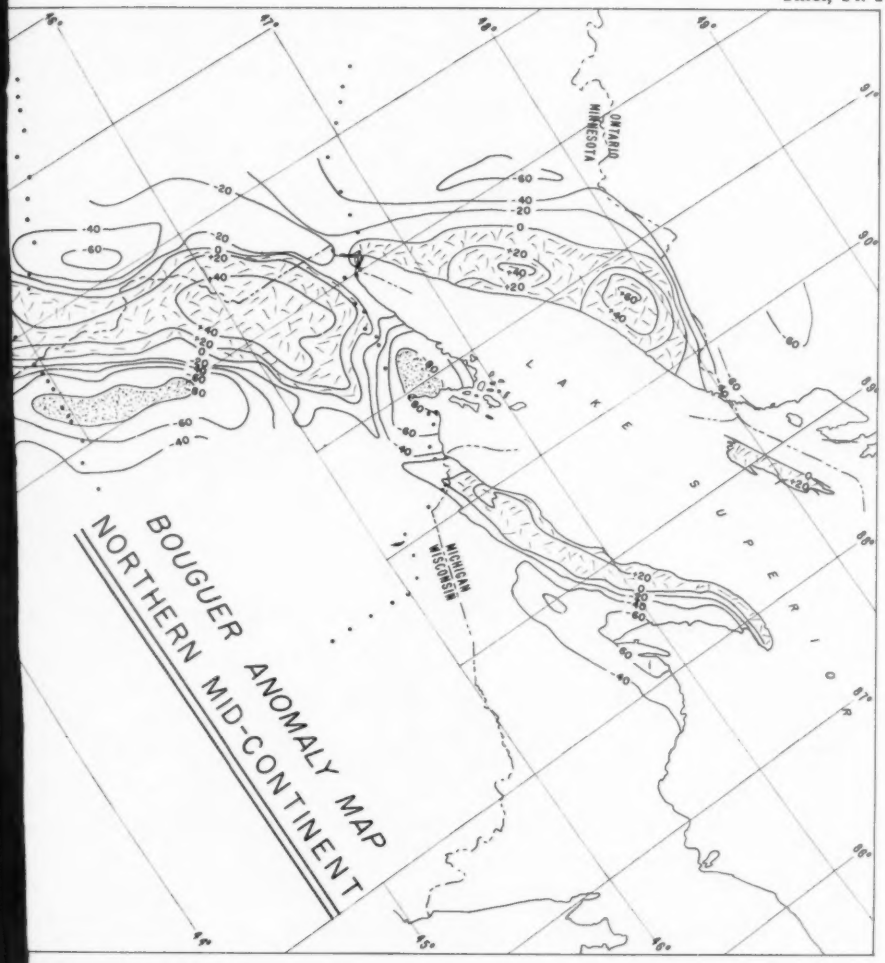
PROFILES



FILES ALONG YOSEMITE VALLEY



BOUGUER ANOMALY MAP OF THE NORTHERN MIDCONTINE



MIDCONTINENT

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CORRELATION OF GRAVITY ANOMALIES WITH THE KEWEENAWAN GEOLOGY OF WISCONSIN AND MINNESOTA

BY EDWARD THIEL

ABSTRACT

Gravitational mapping by the University of Wisconsin has delineated what appears to be the largest positive-anomaly feature on the North American continent, extending from the Lake Superior region southwest into Kansas. For the greater part of its length this "midcontinent gravity high" is flanked on both sides by gravity lows. Because the southern part of the anomalous area is blanketed by Paleozoic sediments, the cause of the anomaly was sought first at its northern end, around Lake Superior, where Precambrian rocks crop out to facilitate a correlation of gravity and geology.

Around western Lake Superior the positive anomalies correlate with Keweenawan lava and gabbro. The negative anomaly on the Bayfield Peninsula reflects a thick accumulation of sandstone and shale which was deposited in the subsiding Lake Superior syncline during Upper Keweenawan time. A second thick accumulation of sedimentary rocks may underlie the gravity low at Cumberland. Steep gravity gradients indicate the Douglas fault. A second major fault symmetric to the Douglas fault is mapped in northwestern Wisconsin on the opposite side of the Lake Superior syncline. The center of the syncline has been thrust upward between the two faults as a horst. A traverse along the spit at Duluth fails to detect the North Shore fault as it is usually mapped; if the fault exists, no great amount of sandstone is placed in juxtaposition to basaltic lava. Detailed correlation of gravity and geology is presented on maps and structure sections along lines of gravity traverse.

The usual isostatic correction cannot reduce the gravity differentials around western Lake Superior. A geological correction which allows for the effects of near-surface geology does account for most of the anomalies. Any attempt to compute the extent of crustal warping at depth without allowing for the near-surface geology would have led to considerable error.

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INTRODUCTION

The Midcontinent Gravity High

In 1949 the Geophysics Section at the University of Wisconsin began a program of regional gravitational and magnetic mapping of the western United States as a means of studying the fundamental geologic structure of the area. Observations also were taken en route to the West and upon return to Madison. The result in the Midwest was a series of traverses with a station interval of 6 to 8 miles along the major highways leading westward from Madison. When the gravity observations were reduced to yield Bouguer anomalies, there was a section having positive Bouguer anomaly values on each traverse. These suggested the existence of a long narrow belt of positive anomalies stretching 800 miles from Lake Superior southward into Kansas. The discovery was announced by Prof. George P. Woollard (1951, p. 636) in an annual report of the Committee for the Geophysical and Geological Study of Continents of the American Geophysical Union.

Plate 1 shows the location of this positive-anomaly feature (referred to below as the "midcontinent gravity high") with its associated flanking lows. The dots indicate the early traverses upon which the existence of this more or less linear feature was postulated. The southernmost traverse through Kansas is part of the first transcontinental gravity traverse established by Woollard (1943) in 1939. The other traverses shown were run by J. C. Rose in 1949.

This gravity-anomaly lineament is one of the most prominent gravitational features in the United States. The local gravitational changes, considering either Bouguer or isostatic anomalies, reach 150 mgals.

Over the southern two-thirds of the anomalous area there is a blanket of Paleozoic and Mesozoic sediments covering the Precambrian basement rocks. The gravity anomalies are of such magnitude that they cannot be accounted for by density variations within this blanket of sediments, and their cause must lie at greater depths. There are only a few boreholes from which samples have been recovered which permit identification of the lithology of the deeper rocks in the area. At the northern end of the anomaly the Precambrian basement rocks crop out, and it is possible to correlate the gravity values with the geology.

This investigation was undertaken to establish the relationship of the gravity anomalies to the lithology and structure of the rocks in the Wisconsin-Minnesota area, since it was believed that there was a direct genetic connection between the gravity values and the geology of the area.

Previous Gravity Studies

In 1917 the United States Coast and Geodetic Survey published the first regional Bouguer gravity–anomaly map of the United States (Bowie, 1917, Fig. 13). This map was based on a limited number of pendulum stations and consequently gave a very generalized picture of the true situation. No pendulum stations were located on the midcontinent gravity high in Kansas, Nebraska, or Iowa, so that this feature was missed in these States. Positive anomalies were observed in southeastern Minnesota at Minneapolis and Faribault; stations at Duluth, Minnesota, and North Tamarack, Michigan, indicated a region of positive anomaly on the southwestern shore of Lake Superior. In the area between Minneapolis and Duluth the scanty control suggested a negative anomaly, and the contour lines were drawn at right angles to their true trend.

A more recent Bouguer anomaly map has been published by Lyons (1950, p. 34–35). This map includes data supplied by State geological surveys, the U. S. Coast and Geodetic Survey, petroleum companies, and individual investigators, and consequently provides a valuable revision of the 1917 map. The most prominent feature of Lyon's map is the positive anomaly extending from Lake Superior to Kansas, although his mapping

differs in detail from that of Plate 1. In general, Lyon's map makes the midcontinent gravity high more linear and continuous than it is in Plate 1. For the area around western Lake Superior his control was the pendulum network of the U. S. Coast and Geodetic Survey.

The only detailed gravity work published for the Keweenawan of Wisconsin and Minnesota prior to the commencement of the work here reported is a traverse across the Douglas fault south of Pine City, Minnesota, by George I. Welch (1941). From the symmetry and magnitude of his gravity profile, Welch concluded that the fault is nearly vertical at this location, with 11,000 feet of sandstone in juxtaposition to basaltic lava.

Acknowledgments

The field mapping of the midcontinent gravity high (Pl. 1) was done by graduate students at the University of Wisconsin under the direction of Prof. G. P. Woollard. W. A. Black is primarily responsible for field observations in the southern part of the anomalous area. In carrying out the field work in Wisconsin and Minnesota the writer was ably assisted by W. R. Bruce, L. F. Machesky, R. E. Randolph, H. J. Richgels, and Monroe Woollard. Others who made independent measurements in the area whose results have been incorporated in this study include J. C. Rose and Donald Plouff.

At the same time, a group under the direction of Prof. Loyal O. Bacon of the Michigan College of Mines and Technology was engaged in gravitational mapping of the Upper Peninsula of Michigan, the third State in which Keweenawan rocks crop out. The geologic interpretation of the gravity data for the Michigan Keweenawan area will be given elsewhere by Professor Bacon.

This paper is based on a thesis submitted in partial fulfillment of the requirements for the degree of Doctor of Philosophy, at the University of Wisconsin, supervised by Professor Woollard. The author wishes particularly to thank Professor Woollard for his encouragement and assistance in all phases of the investigation.

Financial support of the project was provided through a grant by the New Jersey Zinc Company to the University of Wisconsin to promote geophysical studies in the Keweenawan regions of northern Wisconsin.

The author is grateful to the Wisconsin Alumni Research Foundation for a research assistantship covering the early phases of the field work. He also wishes to acknowledge the grant of a graduate fellowship from the National Science Foundation covering part of the period of these investigations.

Conduct of the Survey

Regional Network

The control for the gravitational mapping in Wisconsin and Minnesota consists of a regional station network with observations at about 8-mile intervals along the principal roads, upon which are imposed 10 detailed traverses perpendicular to the strike of the geologic structure, along which observations were made at 0.1- to 0.5-mile intervals. Approximately 2000 observations were taken. The regional stations along the major highways were established by students at the University of Wisconsin at various times using various gravity meters.

It was necessary to reoccupy a number of the regional stations to unify the various groups of data because the calibration constants for the instruments used were furnished with the instruments and were not all established relative to the same standard, because drift control was sometimes poor, and because different base stations were used by different observers. Figure 1 is a plot of the reoccupation differences against observed gravity; the reoccupation values were obtained using Worden gravity meter No. 14b as a standard. Differences obtained for each of the various instruments used were then fitted with the best horizontal straight line, and all observations for a given survey were then adjusted by the amount necessary to bring this "best-fit" line into coincidence with the horizontal axis of the graph. The standard deviation of observed gravity values for the regional stations before adjustment was 1.4 mgals and after adjustment 0.5 mgal. The true estimate of error lies between these two figures, but probably closer to the smaller. Although this

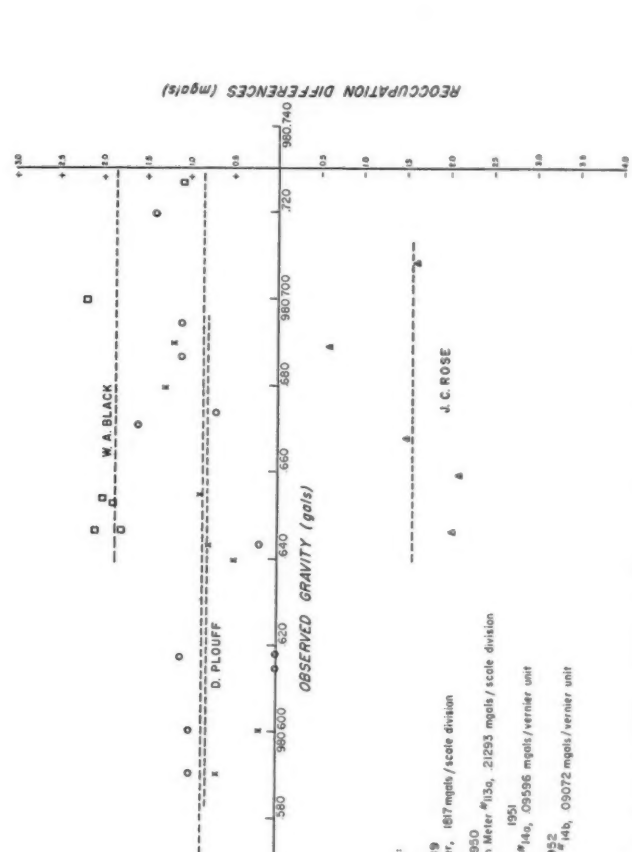


FIGURE 1.—ADJUSTMENT OF THE REGIONAL GRAVITY NETWORK

magnitude of error is unacceptable for precise gravity work, it is considered satisfactory for reconnaissance work where the chief purpose is interpretation of regional geology. The observed gravity values along the detailed traverses taken by the writer have an accuracy believed to be of the order of 0.5 mgal.

TABLE 1.—BASE-STATION DESCRIPTIONS AND VALUES

Madison, Wisconsin—University of Wisconsin campus, basement of Science Hall, on concrete floor immediately outside entrance to Room 7

Observed gravity = 980.3688 gals

Clam Lake, Wisconsin—On the ground, adjacent to east side of post-office porch

Observed gravity = 980.6005 gals

Duluth, Minnesota—St. Louis County Courthouse, Fifth Avenue and Second Street, northeast side of building, on concrete sidewalk at foot of steps

Observed gravity = 980.7613 gals

Detailed Traverses

The seven detailed traverses south of Duluth were run by the author and his assistants during the summer of 1952 (Fig. 3). The station interval is 0.5 mile, except where the Gogebic Iron Range of Wisconsin is crossed; there the interval was reduced to 0.1 mile. Headquarters during the 1952 field season were maintained at Clam Lake, Wisconsin. Each traverse began and closed on Clam Lake. The Madison gravity base station was tied to Clam Lake directly by a closed loop and indirectly at other times by reoccupying stations on several of the traverses.

The three detailed traverses in Minnesota were run during the summer of 1954. The station interval is again 0.5 mile. The gravity work of the 1954 field season is based on a station at the St. Louis County Courthouse in Duluth. During the course of the work in the Lake Superior region, this station was tied to the Madison gravity base station seven times by various observers.

The absolute basis of the present survey rests on the values of the Madison, Clam Lake, and Duluth base stations (Table 1). The Madison value (980.3688) is an average of gravimeter and pendulum ties between Madison and Washington, D.C., assuming 980.1190 gals for the U. S. Coast and Geodetic Survey gravity base at the Commerce Building in Washington.

TABLE 2.—PERTINENT DATA FOR REDUCTION OF THE OBSERVATIONS

Position control, Wisconsin—USGS planimetric maps, 15 minute series

Position control, Minnesota—county highway maps, various scales

Elevation control—railroad elevations and USCGS, USGS, US Lake Survey, and State Geological Survey benchmarks where available; otherwise, the average reading of 4 altimeters

Theoretical sea-level gravity—1930 International Gravity Formula

Assumed density of rocks above sea level—2.67 gm/cc

Elevation and mass correction factor—.06 mgal/ft.

Reduction of Data

Worden gravity meter No. 14b with a linear calibration constant of 9.0725 mgal/100 vernier units was used during the 1952 field season. This calibration was obtained by calibrating No. 14b against the University of Wisconsin geodetic meter, Worden 10e, over a latitude range from Brownsville, Texas, to Jamestown, North Dakota. Meter No. 10e was in turn calibrated against the Cambridge pendulums in Australia and Great Britain.

Worden meter No. 14d was used during the 1954 field season. The tilt-table calibration provided by the manufacturer was 10.327 mgal/100 vernier units. A field calibration of 10.329 mgal/100 vernier units was obtained by running the meter over the University of Wisconsin pendulum calibration line between Great Falls, Montana, and Fairbanks, Alaska. The field calibration was adopted for this work.

All observations were reduced to provide simple Bouguer anomalies. Table 2 summarizes the pertinent information relative to the reduction of the observations. Except as noted, no corrections have been made for departure of topography from the assumed level sheet in the vicinity of the station, or for the gravity

effect of earth tides. For most stations the terrain correction is less than 0.5 mgal; for a few stations on the Gogebic Range and the North Shore it attains a maximum value of 2 mgals. The tidal effect is partially incorporated in the instrument drift correction and cannot in any case exceed 0.3 mgal. Because both effects are small in comparison with the observed anomalies, the expense of making the corrections is not justified from the point of view of added scientific value; the corrections are of the same order of magnitude as the errors resulting from inaccuracies of the barometric leveling.

Heiskanen (1953) argued that Bouguer anomalies can be misleading when used for general geophysical studies, particularly in mountainous areas, along oceanic margins, and on oceanic islands. The argument is that isostatic compensation affects the anomalies through the variation of density in adjacent crustal columns (Pratt-Hayford Theory) or through the existence of low-density roots beneath elevated columns (Airy Theory). Therefore, it is important to determine the effect of an isostatic correction for the area under study.

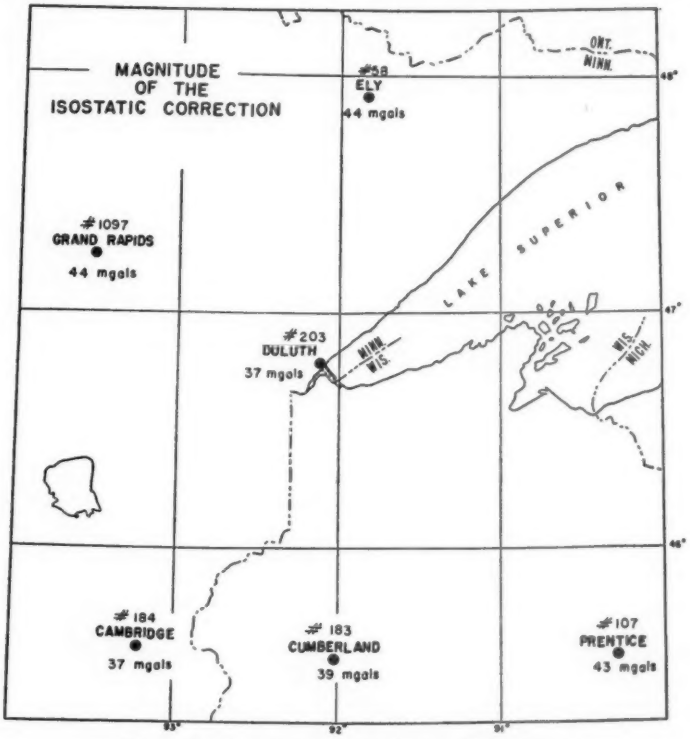


FIGURE 2.—MAGNITUDE OF THE ISOSTATIC CORRECTION

In Figure 2 the U. S. Coast and Geodetic Survey pendulum stations in the area are indicated. The figure beside each station is the number of mgals that must be added to the Bouguer anomaly at the station to convert it to an isostatic anomaly reduced according to the Pratt-Hayford theory, with the assumed depth of compensation at 96 km (Duerksen, 1949). This figure is nearly constant for the area, which is to be expected since the region is not mountainous. Thus, although the isostatic correction would change all anomaly values by 40 mgals, it cannot greatly alter the contour pattern of the map, for the relative

anomaly differences are not reduced by applying an isostatic correction.

The author's Ph.D. thesis, on file in the library of the University of Wisconsin, contains station descriptions and principal facts for the area surveyed. Geophysicists who wish to tie local surveys made for commercial purposes into the existing absolute gravity network are referred to this work.

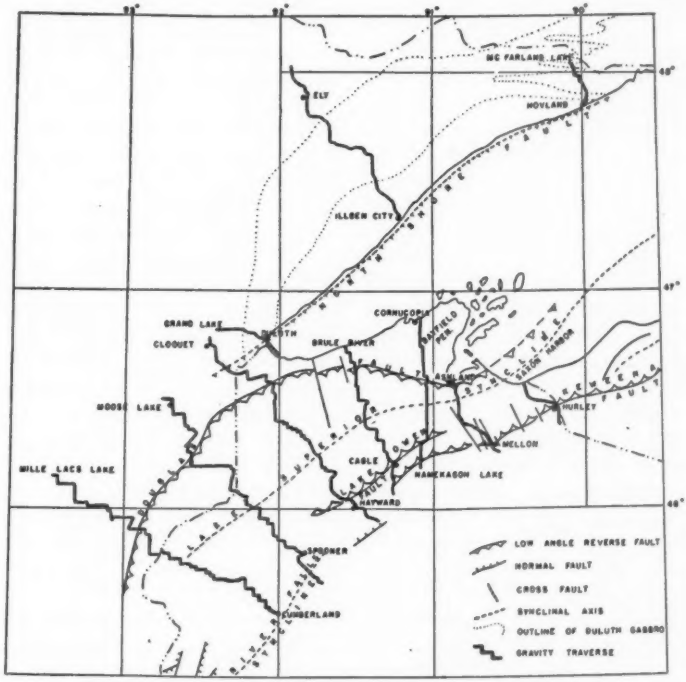


FIGURE 3.—TECTONIC FEATURES AND GRAVITY TRAVERSES
After King *et al.* (1944)

CORRELATION OF GRAVITY AND GEOLOGY

Geology of the Keweenawan System

The Keweenawan series of the Lake Superior region was formed in late Precambrian time. The first event of the period was the continental deposition of the Lower Keweenawan. This sequence is best exhibited on the Canadian shore of Lake Superior, where 1400 feet of clastic sediments have accumulated, although a thin accumulation of Lower Keweenawan conglomerate and quartzite is also recognized on the south shore.

The Middle Keweenawan is characterized by enormous extrusions of igneous rocks (Pl. 2). These are predominantly basic amygdaloidal lava flows. Sandstones and conglomerates are interbedded with the flows and become increasingly abundant upward in the sequence. The widespread extent of the flows and the paucity of volcanic ash suggest that the flows issued from a fissure or system of fissures rather than from volcanoes. Associated with the flows and intruded into them are numerous dikes and sills, dominantly basic. The most prominent is the Duluth gabbro on the northwest shore of Lake Superior.

The period of extrusive eruption was followed by one of predominantly continental deposition with an estimated 25,000 feet of sediments. These deposits consist chiefly of red feldspathic shaly sandstones at the base (Oronto group) exhibiting rapid lateral varia-

tion and irregular bedding. Upward the sandstones acquire a more quartzose character (Bayfield group); Thwaites (1912) considered the transition conformable.

During the evolution of the Middle and Upper Keweenawan a syncline was forming in the area of accumulation. Hotchkiss (1923), working on the south shore, interpreted bent-pipe amygdules, fanning of dips indicating thinning of flows away from their source, and squeezing up of underlying mud in front of the flows as indications that the main source of the lava lay in the area now covered by Lake Superior. He considered that the withdrawal of this great mass of molten material, aided by its tremendous weight after extrusion, caused a collapse of the roof to form a structural basin. This collapse was gradual, so that while extrusion continued the source area maintained its elevation. Thus an increasing diminution is observed today in the dip of the individual lava flows going upward in the sequence. With the cessation of eruption, downwarping continued, and the arkosic clastics of the Upper Keweenawan were deposited in the trough. The younger quartzose sediments dip only slightly inasmuch as they were probably deposited after much of the downwarping had been accomplished. The syncline as a whole is asymmetrical, with the steeper dips occurring on the south shore.

The Keweenawan is broken by several large faults and many small ones (Fig. 3). The Douglas and Keweenaw faults are found on opposite sides of the trough, with thrusting away from the axis. A much younger fault, considered in detail below, has been postulated on physiographic grounds to outline the northwest shore of Lake Superior. Most geologists (Van Hise and Leith, 1911, p. 623; Thwaites, 1935, p. 225) consider the early faulting, and possibly also the downwarping of the trough, a result of lateral compression.

Density Values

Although the Keweenawan of Wisconsin has been studied for over 75 years, few density determinations have been recorded. For the igneous rocks in the area the author used measurements by R. Pumpelly (1880), A. A. Julien (1880), R. D. Irving (1883), and A. F. Gregory, a graduate student at the University of Wisconsin.

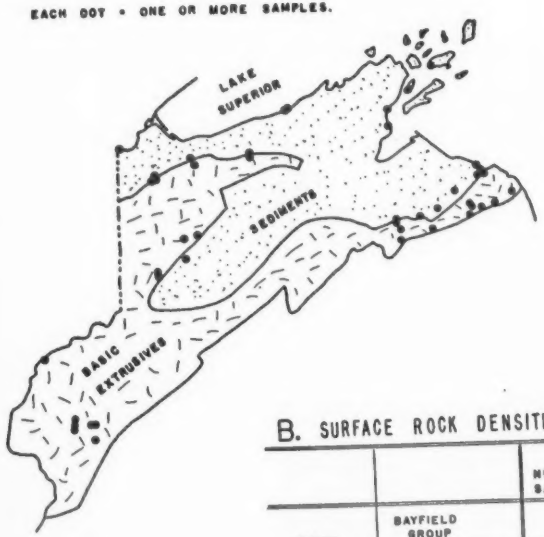
Except for the work of Buckley (1898), few determinations of the density of the sedimentary rocks and none of porosity could be found in the literature. Therefore, a limited number of measurements was made on specimens in the University of Wisconsin rock collection. Mineral densities and porosities were determined by the immersion method of Buckley, in which the dried specimen is weighed in air, and the saturated specimen is weighed in air and in water. The saturation is accomplished by gradual immersion under reduced pressure. Bulk densities are then computed. As a check, bulk densities were also determined by the immersion method of Melcher (1921), in which the dried specimen is weighed in air, then coated with paraffin and weighed in water. Areas of sampling, number of determinations available, and adopted surface-rock densities are indicated in Figure 4A, B.

The question of how density varies with depth is difficult to assess, particularly in the case of the clastic sedimentary rocks. Several factors must be considered: (1) increase of temperature causing the rock to expand; (2) increase of pressure compressing the rock; (3) combined effects of temperature and pressure on the density of the interstitial water; (4) change of porosity with depth, including the question of whether pore space can be maintained at the depths considered; (5) cementation; (6) changes in the degree of saturation of porous rocks with depth; (7) metamorphic changes due to heat and pressure.

The cubical coefficient of thermal expansion of sandstone is 30×10^{-6} per centigrade degree (Birch *et al.*, 1942, p. 37). Using the mean of Van Orstrand's limits for the geothermal gradient of sedimentary rocks (Van Orstrand, 1951, p. 141), 7.15° C per 1000 feet, the decrease of density at 25,000 feet is .012. The pressure at this depth is about 1770 bars, which, with a compressibility of 25.6×10^{-7} reciprocal bars (Birch *et al.*, 1942, p. 61), gives a density increase of .011. These two effects, which are small for sandstone, work in opposite directions so that no great error is introduced. Calculations also indicate that the combined

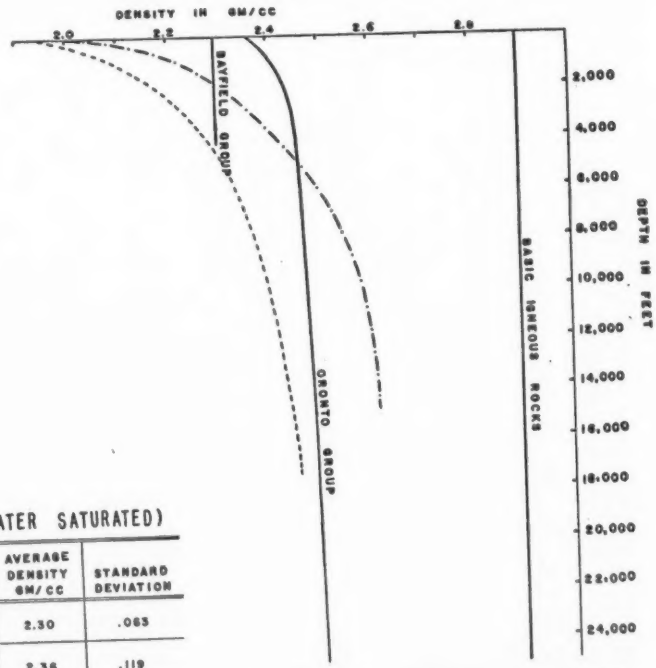
A. DISTRIBUTION OF SAMPLES

EACH DOT = ONE OR MORE SAMPLES.



B. SURFACE ROCK DENSITIES (WATER SATURATED)

		NO. OF SAMPLES	AVERAGE DENSITY GM/CC	STANDARD DEVIATION
UPPER KEWEENAWAN	BAYFIELD GROUP	20	2.30	.083
	ORONTO GROUP	10	2.36	.119
MIDDLE KEWEENAWAN	BASIC IGNEOUS ROCKS	50	2.90	.097
LOWER KEWEENAWAN	QUARTZITE & CONGLOMERATE	THICKNESS < 100 FEET; NO VALUES.		



C. DENSITY VARIATION WITH DEPTH

SOLID LINES: WISCONSIN KEWEENAWAN

DASHED LINES: GULF COAST SEDIMENTS
---- AFTER NETTLETON
-·-· AFTER PETERS & DUGAN

FIGURE 4.—ROCK-DENSITY DATA

effect of temperature and pressure upon the interstitial water is not likely to be significant at the maximum depth considered.

The density of argillaceous rocks, however, is greatly affected by pressure. The work of Athy (1930) indicates an exponential increase in density of shales with depth. From Thwaites' (1912) detailed descriptions of the Upper Keweenawan sediments, it is possible to estimate the relative amounts of sandstone and shale present. The following values were adopted: Oronto group, 40% sandstone, 60% shale; Bayfield group, 99% sandstone, 1% shale.

The possibility that at some depth the pressure will be sufficiently great to close all pore space was considered by Hoskins (1896). He concluded that permanent cavities occupied by water under hydrostatic pressure would close at a depth of 10,350 m. As this is a greater depth than is believed to obtain in the Lake Superior syncline, it appears safe to ignore this possibility.

Knowledge of the effect of cementation on porosity has been extended recently by the experimental work of Maxwell and Verrall (1954). For a pure quartz sand they find that no cementation occurs for pressures equivalent to depths of less than 24,000 feet and at temperatures below 270°C. If the sand is subjected to greater depths and temperatures, appreciable porosity (>10 per cent) will not be retained. For an impure sandstone the reduction of porosity is greater than for a quartz sand under similar conditions, and the limiting depth for appreciable porosity is thus less.

The question of the degree of saturation with depth embraces a wide divergence of opinion, some authors holding that saturation decreases with depth (Fuller, 1906), while others consider it more likely that the pores are filled (Birch *et al.*, 1942, p. 17; Evans and Crompton, 1946, p. 219). For the Upper Keweenawan rocks, which are to a large extent porous sandstone, it appears reasonable to assume complete saturation at depth.

Metamorphic changes caused by heat and pressure probably are not significant in the Keweenawan rocks, as they are not greatly metamorphosed nor orogenically deformed to the same extent as the underlying Huronian.

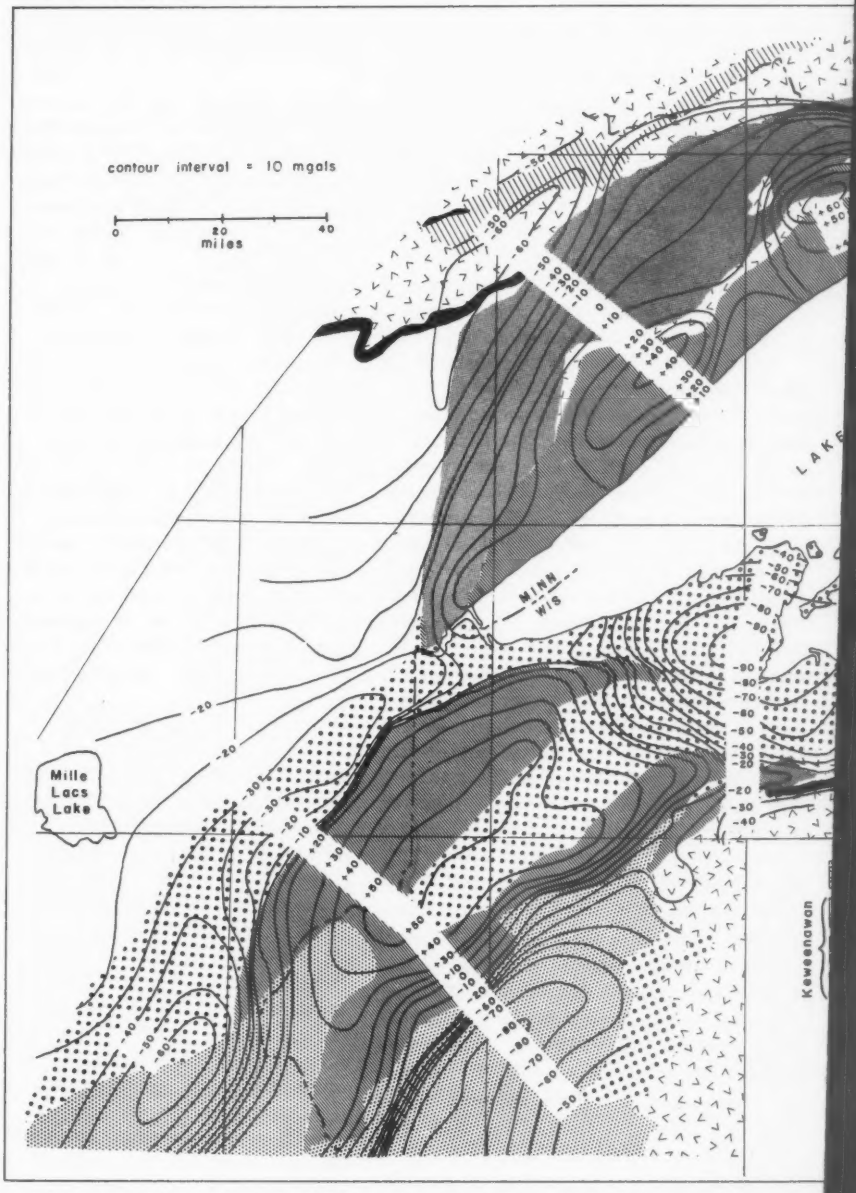
On the basis of the above considerations the density variations with depth indicated in Figure 4C were adopted. For purposes of comparison, two density-depth curves for the Gulf Coast are shown, one used by Nettleton (1934) in his study of the mechanics of salt-dome formation and the other by Peters and Dugan (1948) for their gravity study of the Grand Saline salt dome. The two areas are geologically dissimilar, but it is of interest to compare the assumptions made in the Lake Superior area with data based on extensive geophysical work and deep drilling in another area. Variation of density values with depth is the greatest source of uncertainty in the present study.

Gravity Map

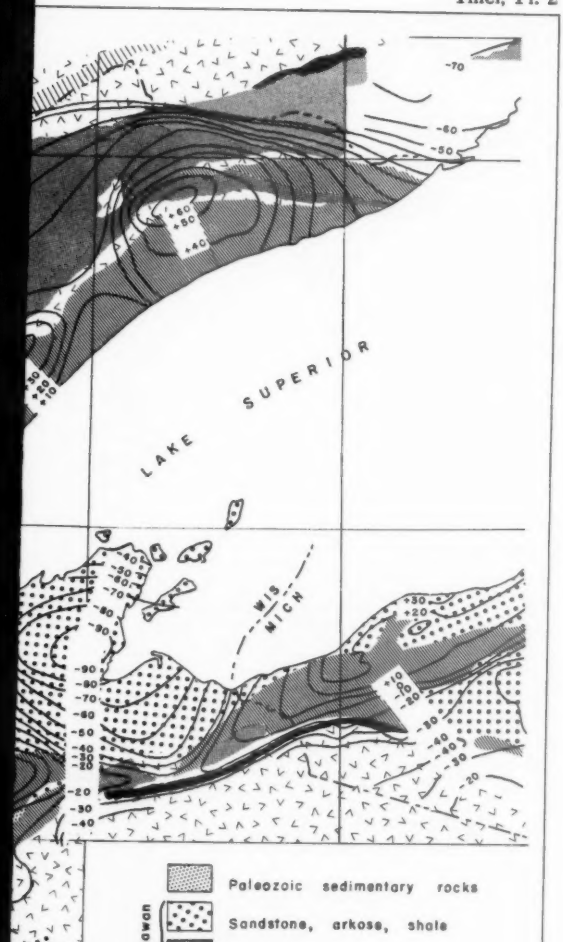
Plate 2 presents the regional gravity map of the Wisconsin-Minnesota Keweenawan area superimposed on the generalized U. S. Geological Survey map of the Lake Superior region (Leith, Lund, and Leith, 1935). As noted previously, the expected value of the Bouguer anomaly for this region is not 0 but, because of isostasy, about −40 mgals. Regions of the map which depart considerably from this value indicate unusual conditions which deserve investigation.

The correlation of gravity and geology exhibited by Plate 2 is striking. The positive anomalies occur over the dense basic rocks, either basaltic lava flows or gabbro. The negative anomaly over the Bayfield Peninsula is associated with the downwarped Lake Superior syncline which has been filled with low-density sandstones and shales of the Upper Keweenawan.

The gravity low beginning at Hayward and extending through Cumberland is more puzzling. This area is heavily mantled by glacial drift with less than 1 per cent of the surface exposed as outcrop and with no deep wells to guide geologic inference. On the basis of the gravity observations, it is suggested that this low also reflects a thick accumulation of Upper Keweenawan sediments. The River Falls syncline, exhibited by the exposed Paleozoic rocks, might thus indicate a more significant structure in the Keweenawan sediments at



GRAVITY AND GEOLOGY OF THE WESTERN LAK



ERN LAKE SUPERIOR REGION

depth. The tentative identification by Thwaites (Personal communication) of Upper Keweenawan sediments in a well at Hayward supports this view.

The extension of the positive anomaly toward the Keweenaw Peninsula is interrupted in the vicinity of Mellen (Pl. 1; Pl. 2). The reduced gravitational attraction there is probably caused by the intrusion of a large mass of lower-density granite as mapped by Aldrich (1929, Pl. 1).

The high anomaly gradients bounding the lava mass suggest faults between high-density lava and low-density sandstone and shale. The tectonic map (Fig. 3) shows that for the northwestern boundary of the flows this inference is valid; the closely spaced contours accurately locate the Douglas fault. The high gradient to the southeast suggests a second major fault symmetric to the Douglas fault. Under this interpretation the axial portion of the Keweenawan trough in Douglas County has been thrust upward as a horst. The fault suggested on the southeast by the data is probably the extension of the Lake Owen thrust mapped by Aldrich (1929, Pl. 1) and is here referred to by that name.

Gravity Traverses

Plate 3 presents a more detailed correlation of gravity and geology. Structure sections have been drawn for the six longest traverses, with locations as indicated on Figure 3. In drawing the sections, geologic maps by Thwaites (1912; 1935), Hotchkiss (1915), Hotchkiss and Bean (1929), Aldrich (1929), Leith, Lund, and Leith (1935), Grout *et al.* (1932) and the Wisconsin Geological and Natural History Survey (1939) were utilized. Except as noted, the geology is not "fitted" to the gravity curve; rather, the observed anomaly values are correlated with the geology as set forth in the literature.

The Namekagon Lake-Cornucopia traverse (Pl. 3, Section A) begins on County Highway D south of Namekagon Lake and runs northward across the Gogebic Range, through the town of Grandview, and across the Bayfield Peninsula, terminating northeast of Cornucopia on Squaw Point. The traverse begins at its southern end on inferred Huronian sediments with the average regional value of −40 mgals. The anomalies increase as the basic lava is approached, with a sharp 4 mgal rise upon crossing the Ironwood iron formation. A small dip in the curve occurs where conglomerate has been wedged between flows as a result of faulting. The greatest deficiency of gravity for the entire area surveyed was discovered on this traverse. The curve dips to −92 mgals on the Bayfield Peninsula west of Washburn. The accumulation of sediments is inferred to be maximum here. The anomalies increase as Cornucopia is approached, suggesting that the underlying lava is approaching the surface.

The Cable-Brule River traverse (Section B) begins north of Twin Lakes on State Highway 77 and runs northward through Cable, passing between the towns of Brule and Iron River and terminating on Lake Superior at the mouth of the Brule River. The traverse begins at the south on inferred Huronian sediments. The gravity values increase over the flows, drop 8 mgals over a wedge of sandstone between flows, and then reach a maximum. The Lake Owen thrust is clearly indicated. The fact that the gravity values do not drop as low as those farther east suggests that the sedimentary section is much thinner here. Where the lava again reaches the surface, the profile rises. Upon crossing the Douglas fault there is a sharp drop in gravity, after which the profile rises, suggesting that the underlying flows once more are approaching the surface.

The Hayward-Cloquet traverse (Section C) begins on the east side of Grindstone Lake, passes through the towns of Hayward, Gordon, and Cloquet, terminating 3 miles north of Cloquet. This traverse exhibits even more clearly the strong positive anomalies over the dense amygdaloidal lava flows. The decrease in the thickness of the sandstone in the syncline should be noted. Since this profile across the lava mass is the most symmetric and least disturbed by the sedimentary accumulations adjacent to the faults, it was used to calculate the thickness of the lava pile in Douglas County. Assuming that the lava mass can be approximated by a rectangular block with a density of 2.90 and that the density of the pre-Kewee-

nawan rocks is 2.67, the thickness of the lava at this location is 33,000 feet.

The Spooner-Moose Lake traverse (Section D) begins on the west side of Long Lake, passes

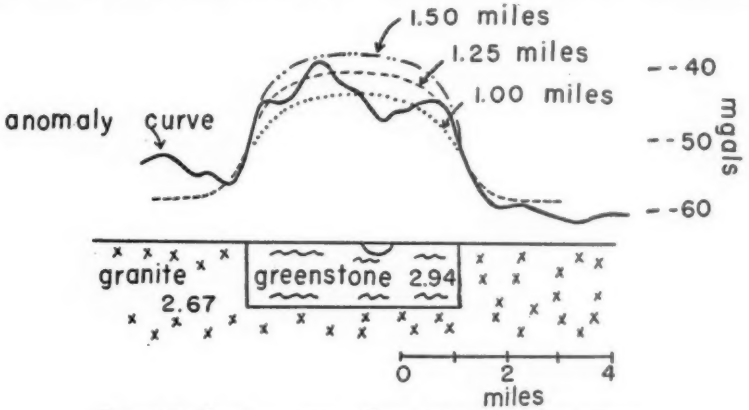


FIGURE 5.—DETERMINATION OF DEPTH OF THE ELY GREENSTONE

through the towns of Spooner, Bruno, and Moose Lake, terminating 6 miles northwest of Moose Lake. The Cumberland-Mille Lacs Lake traverse (Section E) originates at Cumberland and runs northwest, crosses the St. Croix River just south of its junction with the Snake River, passes through Brook Park, and terminates at Isle on the shore of Mille Lacs Lake. Both traverses cross the structure perpendicular to the strike. The steep gradients over the Douglas fault on the west and the Lake Owen fault on the east are some of the most pronounced changes of gravity observed in the country. Arrows indicate the present limit of the westward extent of Cambrian sediments. The gravity profiles show that the lavas extend for some distance eastward below this thin cover of Cambrian sandstone and glacial drift.

The Illgen City—Ely traverse (Section F) crosses the Keweenawan section of northeastern Minnesota. It begins at the junction of U. S. Highway 61 and Minnesota State Highway 1, following the latter through Finland and Isabella toward Ely, and terminating about 10 miles north of Ely on the Echo Trail. A large positive anomaly is again observed over the basic lava, with a rather sharp decrease upon crossing the lower-density red-rock phase of the Duluth gabbro. A second prominent anomaly occurs over altered basic lavas of Archean type near Ely. Figure 5 shows an attempt to compute the thickness of these older lavas. A rectangular block of infinite extent in the direction perpendicular to the traverse has been used to approximate the altered lavas, and theoretical curves have been computed for various assumed depths. A density of 2.67 is assumed for the granite country rock. The greenstone density value of 2.94 is an average of 27 determinations furnished by R. H. Chapman. Although the observed anomaly curve is somewhat irregular (the low near its center correlates with infolded slates of Knife Lake age), it is concluded that the base of the greenstone lies at a depth of 1 to 1½ miles.

Geological Correction

The excellence of the above qualitative correlation of gravity and geology is impressive, but whether the near-surface geology will quantitatively account for the anomalies or whether they reflect in part the effects of crustal warping at greater depths must still be determined.

Many geologists have mentioned the importance of allowing for local geology in gravity studies, but few have undertaken quantitative calculation. The first reasonably complete qualitative discussion was given by David

White (1924). He briefly described the geology around locations where gravity had been measured with the pendulum and commented upon the gravitational effects that this geology might exert. The qualitative correlation of gravity and geology so impressed White that as one of his principal conclusions he wrote:

"The inspection of the anomalies, with brief examination of the local geologic data, encourages the view that the rock materials of excessive or deficient weight causing the principal portion of the anomalies lie relatively near the surface, and that allowance in the computations for such matter would largely liquidate the anomalies."

With respect to the Precambrian region about western Lake Superior he said:

"The demonstrated presence in portions of Wisconsin, upper Michigan, and Minnesota of synclines and blocks of very thick Precambrian formations, some of which carry heavy igneous rocks or great deposits of iron ore, and some of which are below the average in density, offers exceptional opportunities to observe the possible effects on the pendulum of unusually heavy masses and sheets of rock comparatively near the surface of the ground. . . . In this study regard must be paid to the extent, shape, and attitude of the heavy rock masses, as well as to their depth, thickness, and relations or structure."

However, White did not undertake such computations.

The first quantitative discussion of the subject dealt with the Black Hills-Bighorn-Beartooth region. Chamberlin (1935) and Woollard (1938) computed geological corrections. Their methods produced anomalies which " . . . lump together two important variables—the geology on the one hand, and the isostatic compensation, on the other. So long as one remains a variable, the other cannot be determined" (Chamberlin, 1935, p. 407). Chamberlin then recommended, but did not apply, a method which would consider the geological correction prior to the correction for isostasy.

Perhaps the most significant study of the effect of local geology upon the gravity anomalies is that of Evans and Crompton (1946). These investigators computed geological corrections for an area in Burma and northeastern India where the Tertiary sediments reach thicknesses of over 50,000 feet and where over 6000 gravity observations were available.

Although White felt that anomalies could be compensated by abnormal densities at depths less than 50,000 feet, quantitative calculations have not supported this view. For the Black Hills-Bighorn-Beartooth region, where the qualitative correlation is most striking, the geological correction can account for only a part of the observed anomalies, and large "geo-isostatic anomalies" remain (Woollard, 1938, Fig. 2). The same conclusion was reached by Evans and Crompton:

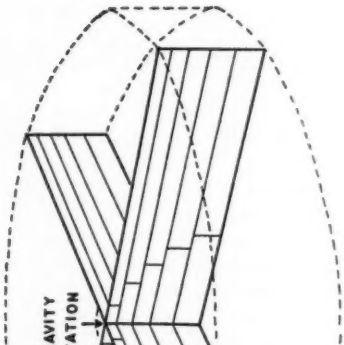
"The work has shown that the suggestion that isostatic anomalies could be accounted for by local geological features is not applicable in eastern India and Burma. After allowance for the effects of known or inferred geology there remain geologically corrected Bouguer anomalies which in extreme cases exceed 175 milligals; even when corrected for isostatic compensation the resulting geologically corrected isostatic anomalies may reach 100 milligals."

The only safe way to determine whether or not the anomalies of northwestern Wisconsin resulted entirely from near-surface geology was quantitative analysis. Figure 6 illustrates the computation method of Evans and Crompton, used in the present analysis. Two cross sections are drawn at right angles through each gravity station. These are superimposed on a template of 20 rectangular compartments, so adjusted in size that the rocks within each compartment produce an anomaly of 1 mgal for each 0.1 gm/cc variation in density. The accepted average density from which variations are measured is 2.67 gm/cc.

The geological correction defined by this procedure takes account of the geology within a radius of 20 miles around the station and to a depth of 11⅓ km (38,000 feet) below it. It is not claimed that the geology is accurately known to such a great depth, but a shallower depth would exclude features whose effects are important. The structure sections of Plate 3 were among those used in the computation.

Geological corrections were computed at 28 stations, and the geologically corrected Bouguer anomalies are indicated in Figure 7. That the Bouguer anomalies are greatly reduced by the application of the geological correction is evident. Whereas the original Bouguer anomaly map displayed a maximum gravity differential of 150 mgals, this is reduced to 40 mgals by the geological correction, a reduction of nearly 75 per cent.

One disturbing feature remains. Although the



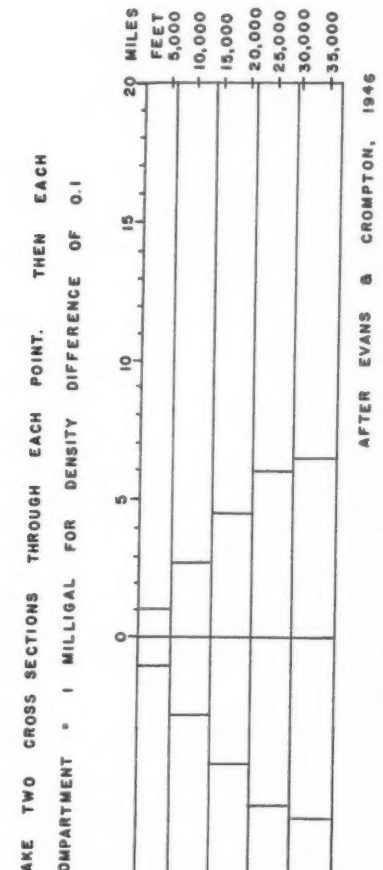


FIGURE 6.—CHART FOR DETERMINING GEOLOGICAL CORRECTIONS TO BOUGUER ANOMALIES

effect of the lavas is largely eliminated, a residual effect still appears over the Bayfield Peninsula and at Cumberland, where the sedimentary accumulations are inferred to be thickest. There are several possible explanations: (1) the sedimentary sections may be even deeper than indicated in Plate 3; (2) the density of the sediments may not increase with depth as rapidly as indicated in Figure 4; (3) there may be a contribution from depths deeper than here considered.

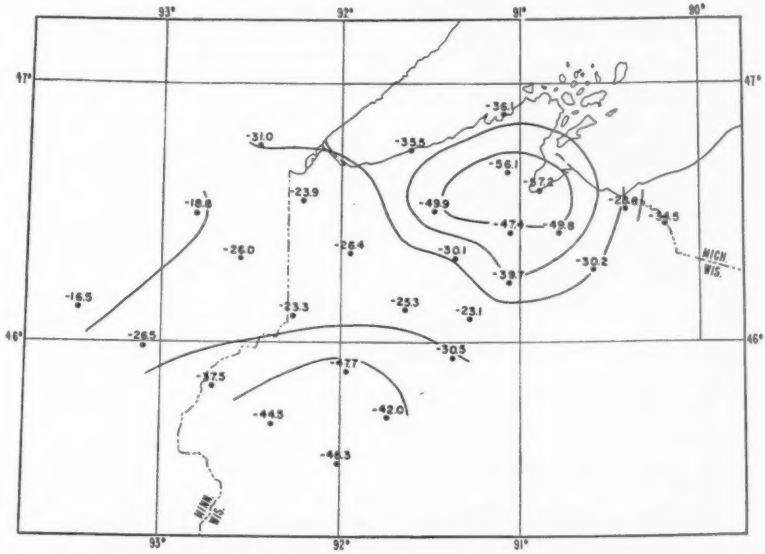


FIGURE 7.—BOUGUER ANOMALIES AFTER APPLYING A GEOLOGICAL CORRECTION

In connection with (3), Plate 1 shows a region of excess mass, flanked on both sides by regions of mass deficiency, which might be caused by crustal warping (Glennie, 1932), by a type of regional isostatic compensation (Vening Meinesz *et al.*, 1934), or by a deformed elastic crust under a superimposed load (Gunn, 1949). According to the latter two proposals, an excess mass such as the Keweenawan lavas would cause a regional downwarp of the crust. Low-density sediments might then be washed into the adjacent flanking areas. Part of the observed anomalies would be caused by high- and low-density material in juxtaposition, but part would be the result of the displacement of heavier subcrustal material by the downwarped crust. Gunn has shown how, under his theory of isobaric equilibrium, an area can be tectonically stable and still exhibit large gravity anomalies. This aspect of the problem is being undertaken by Mr. W. A. Black for the complete mid-continent gravity high and will not be considered here.

LOCAL PROBLEMS OF KEWEENAWAN GEOLOGY

Extension of the Douglas Fault

The most easterly location at which the Douglas fault is exposed is 10 miles west of Brule, where the Middle River crosses the contact of the flows with sandstones of the Orienta group. From this point the fault has been traced eastward by magnetic methods and by inferences based upon discontinuous outcrops. On the Brule River, for example, the actual contact is not exposed but must run between outcrops of lava and sandstone about a quarter of a mile apart. The most easterly outcrop of the flows occurs on Iron River near the western boundary of Bayfield County. On the basis of dip-needle measurements by the Wisconsin Geological Survey, the contact has been traced eastward for about a mile.

In his original work on the Keweenawan, Thwaites (1912) states: "The Douglas fault does not pass through the Apostle Islands but appears to die out into a fold which in turn becomes gentler and finally flattens out beneath the drift-covered area of the Bayfield ridge." However, he later suggested (1935) that the fault curves southeast, passing south of Chequamegon Bay and striking into Lake Superior at a point between Ashland and the Michigan border. The latter interpretation was accepted by the editors of the Tectonic Map of United States (King *et al.*, 1944) (Fig. 3).

Several of the detailed gravity traverses cross this inferred extension of the Douglas fault. On the Cable-Brule River traverse the fault is definitely reflected in the gravity observations. The structure section for the Namekagon Lake-Cornucopia traverse is taken from Thwaites' 1912 map; the possible fault there indicated would be the extension of the Douglas fault on the newer view. The rapid decrease of anomaly values to the north reflects the northward dip of the sandstone-lava contact, but superimposed on this regional trend there is a broad low-amplitude warp of the curve which may indicate a fault or fold at depth. This warp lies north of the fault's position as mapped by Thwaites.

Farther east, near Ashland, there is exposed a large mass of lava surrounded by the clay plain. In 1912 Thwaites concluded that this was an enormous boulder, as no rock was found in neighboring wells.

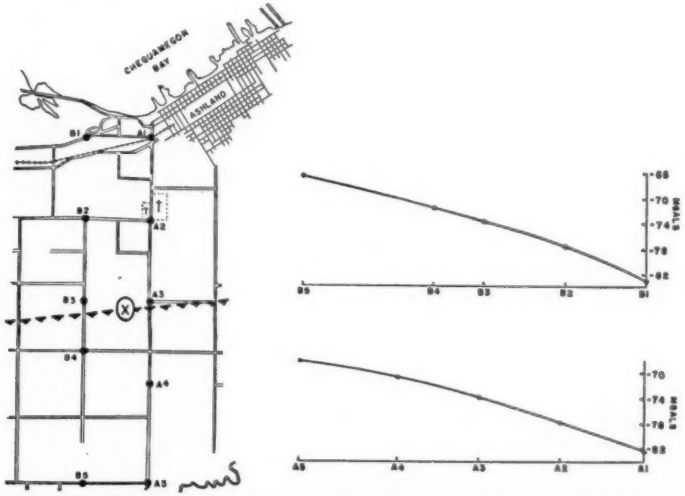


FIGURE 8.—GRAVITY TRAVERSES NEAR TRAP OCCURRENCE SOUTH OF ASHLAND

Two gravity traverses were run past the lava mass (Fig. 8). Both show a steady trend consistent with a general northward dip of the underlying lava-sandstone contact, rather than the steep gradients which had been observed over the Douglas fault farther west. These observations suggest that the mass of lava is not an outcrop. If the Douglas fault does pass beneath this region, the throw is so small or the depth to the highest point on the lava-sandstone contact so great that gravity measurements at the surface can no longer distinguish the fault from a uniformly dipping contact. Also, the fault may lie just north of Ashland, in the area covered by the waters of Chequamegon Bay.

Depth of Drift on the Bayfield Peninsula

The Bayfield Peninsula is a broad ridge stretching into Lake Superior. Irving (1880, p. 14) concluded that it was a continuation of the lava flows which crop out to the southwest: "The central ridge of the Bayfield Peninsula

appears to be made up beyond doubt of some of the crystalline rocks of the Keweenaw system." Thwaites, however, argued that "We can clearly discern that the Bayfield Peninsula must owe its origin primarily to different causes than the trap ridge of Keweenaw Point, which it resembles in outline." He attributed its existence to differential erosion (1912, p. 98): "The Bayfield Peninsula, then, is a ridge of flat-lying, comparatively resistant sandstone, left as an erosion remnant between two areas mainly composed of shales and soft arkose."

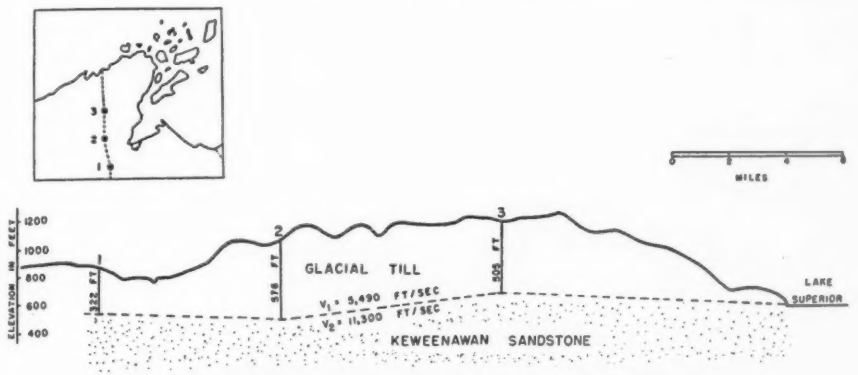


FIGURE 9.—SEISMIC DETERMINATION OF DEPTH OF TILL ON BAYFIELD PENINSULA

During the glacial period a huge interlobate moraine was deposited on the resistant ridge, giving its summit an average elevation of nearly 700 feet above Lake Superior. Except along the shores and some of the deepest valleys there are no exposures of solid rock; that which is exposed is sandstone.

In order to determine whether the large hills on the Bayfield Peninsula are rock-controlled and to establish definitely the nature of the underlying bedrock, several shallow seismic-refraction shots were made on the ridge (Fig. 9). The depth of drift approaches 600 feet. The velocity of the underlying bedrock is 11,300 feet per second and correlates with the measured velocity of sandstone in other localities; it is too low for basalt. These results indicate that (1) the surface topography on the Bayfield Peninsula is erosional, not bedrock-controlled, and (2) the bedrock is sandstone, as stated by Thwaites.

Dip of the Douglas Fault

The Douglas fault is of the reverse type, as indicated by the preceding structure sections. Good exposures occur on the Amnicon River, where the dip of the fault is 45°S. On the Black River farther west the dip of the main fault is about 50°S. (Thwaites, 1912, p. 80).

Figure 10 illustrates an attempt to determine the dip of the fault where the contact is hidden. Part A shows the response expected from three different orientations of the fault plane. Part B shows the locations of the gravity stations, the outcrops of trap and sandstone, and the inferred location of the fault. The observed Bouguer anomaly curve is indicated in Part C. A simple way of determining to which of the three cases the observed profile corresponds is to rotate the observed curve 180° about an axis perpendicular to the plane of the drawing. Comparing C with A shows that the fault at this locality corresponds to case 3—a reverse fault. Using Garland's (1950, p. 9) method of analysis, a dip of approximately 65° is obtained.

From the gravity difference between the asymptotic values of the curve on opposite sides of the fault (in this case 31 mgals), the depth of the sandstone to the north may be calculated, and a value of 4600 feet is obtained. This is considerably less than indicated for this locality on the Cable-Brule River structure section, which is based on the sections published by Thwaites (1912). Thwaites' estimate

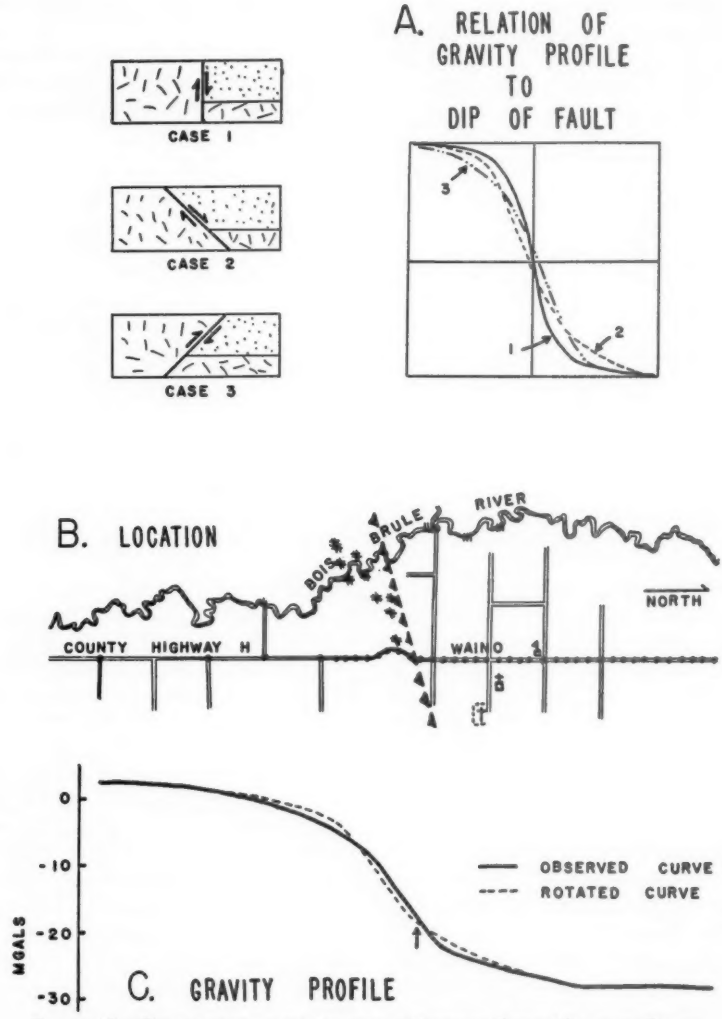


FIGURE 10.—DETERMINATION OF DIP OF THE DOUGLAS FAULT NORTH OF BRULE

of the thickness of sandstone at this point is probably excessive.

Existence of the North Shore Fault

Over 40 years ago Martin (1911, p. 112–115) postulated, wholly on physiographic grounds, that a major fault outlined the northwest shore of Lake Superior. Since then, this fault has been drawn on every regional map of the Lake Superior region, including the Tectonic Map of the United States (King *et al.*, 1944). The concept has been questioned by Minnesota geologists. Schwartz (1949) writes:

> It has been generally assumed that the north shore of Lake Superior and the rock bluff overlooking the St. Louis River is a fault scarp. There is no direct evidence of the existence of the fault. . . ." (p. 81) "The Minnesota coast of Lake Superior in a general way is remarkably regular. This has been suggested as evidence of a fault, but such a conclusion, at least in so far as it is based on topography, is very dubious. . . ." (p. 5)

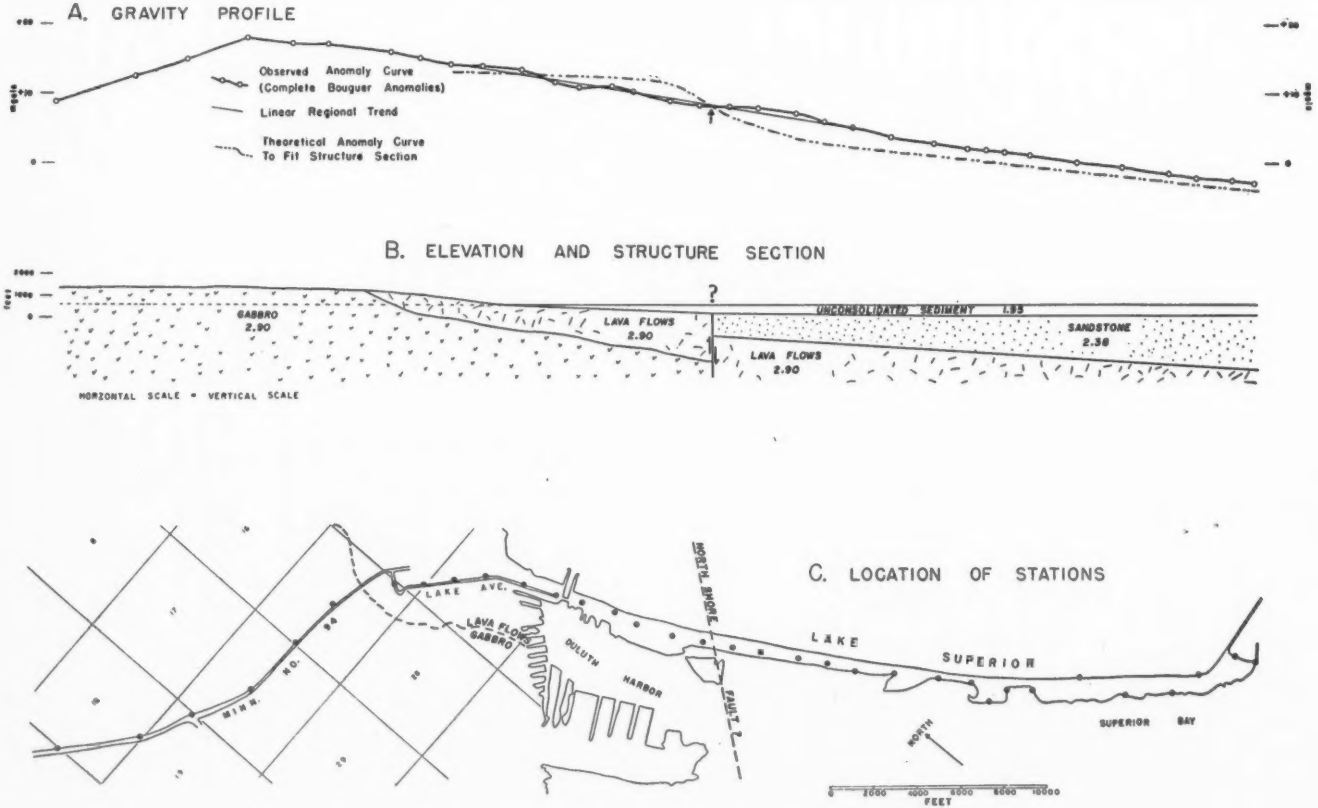


FIGURE 11.—GRAVITY TRAVERSE ACROSS THE HYPOTHETICAL NORTH SHORE FAULT

At Duluth there is an opportunity to contribute to the solution of this question using gravity measurements. Minnesota Point, a spit, extends 7 miles from the north shore. It is separated by only a narrow channel from Wisconsin Point, a similar projection from the south shore. Behind these two spits lies the harbor of Superior-Duluth.

Part C of Figure 11 shows the gravity stations that have been established on Minnesota Point and up the steep bluff through metropolitan Duluth. The North Shore fault is drawn as it usually appears on geological maps. Part B is a hypothetical elevation and structure section, used in computing the theoretical curve which would be obtained if the fault existed with 1000 feet of sandstone in juxtaposition to basaltic lava.

Part A compares the theoretical curve with the observed anomaly curve. A terrain correction has been applied to the observations from curves of terrain effect supplied by Hubbert (1948); the results were checked at three points, using Hammer's tables (1939), with satisfactory agreement.

The three points at the extreme left of the observed curve reflect the approach toward the base of the gabbro, which is in contact with lower-density slates to the west. Except for these points, the observed curve departs, at most, 1½ mgals from a linear regional trend and does not exhibit the distortion expected from a situation corresponding to the hypothetical structure section. The gravity data is consistent with the assumption of a nearly uniform dip of the igneous-sedimentary contact. If the irregularities of the observed curve reflect faulting, the throw is such that no more than 350 feet of sandstone is placed in juxtaposition to basaltic lava.

The gravity traverse does not rule out the possibility of lava faulted against lava, followed by erosion of the fault scarp. The only gravity effect for such a situation would arise from density contrasts at the base of the very thick Middle Keweenawan section. This effect would be spread over a great distance at the surface and could not be separated from the regional trend. The fault would remain undetected.

Basal Dip of the Duluth Gabbro

The Duluth gabbro is one of the greatest basic intrusions known. It crops out in the shape of a great crescent with one end at Duluth and the other in northeastern Minnesota near the International Boundary.

In determining the thickness and size of this great mass, some estimate must be made of its dip. In the past this has been done chiefly on the basis of internal structures; the average dip of the internal banding is assumed to be a good index to the general structure. Grout and Schwartz (1939, p. 14) state that probably "... the dip gradually flattens from an average of at least 25° near the base of the gabbro to an average near Lake Superior of certainly not less than 10°." In a later publication Schwartz (1944, p. 225) states that "The dip is most pronounced at the base where probably 15 to 20 degrees is a fair average."

The dip of the gabbro near its base may be computed from the gravity values, as an example of an independent approach. The Illgen City-Ely traverse (Pl. 3) crosses the base of the gabbro north of Birch Lake, where it is in contact with the underlying Giants Range granite. The rate of increase of gravity near the contact is a measure of the dip of the contact and of the density differential between gabbro (2.90) and granite (2.67). At this locality the observed gradient is 2.7 mgal/km, which gives a basal dip of 17°. The shorter traverse through Duluth (Fig. 3) exhibits a larger gradient near the contact, indicating a greater basal dip in that region.

Summary and Conclusions

The principal results of this investigation may be summarized as follows:

(1) A generalized gravity map has been presented for what appears to be the largest positive-anomaly feature on the North American continent. This midcontinent gravity high has an average width of 30 miles and an amplitude of 100 mgals above the regional gravity value. It extends, with interruptions and offsets, from the Lake Superior region southwest into the Salina Basin of Kansas. For the greater part of its length the high is flanked on both sides by gravity lows.

(2) Around western Lake Superior, where Precambrian rocks are exposed, the positive anomalies correlate with basic lava and gabbro of Keweenawan age. The principal gravity low over the Bayfield Peninsula of Wisconsin correlates with a thick accumulation of Keweenawan sandstone and shale. A similar thick accumulation of sediments may underlie the gravity low beginning at Hayward and extending through Cumberland.

(3) The application of the usual isostatic correction cannot reduce the large gravity differentials around western Lake Superior. A quantitative geological correction does account for the major part of the anomalies; nevertheless, the negative anomalies over the sedimentary basins are not entirely removed.

(4) The gravity work suggests the following comments and possible revisions with regard to the present Tectonic Map of the United States (Fig. 3):

(a) A gravity traverse along Minnesota Point and through metropolitan Duluth fails to substantiate the existence of the North Shore fault as usually mapped (Fig. 11). If the fault exists, its throw is such that no great amount of sediment is placed in juxtaposition to basaltic lava.

(b) Gravity traverses south of Ashland are consistent with a regional dip of the sandstone-lava interface; they provide no evidence for the extension of the Douglas fault south of Ashland. The trap occurrence usually cited as evidence for this fault extension is a boulder rather than an outcrop. If a sandstone-lava fault contact exists in this region, it must be deeply buried.

(c) A steep gravity gradient is mapped to the southeast of the axis of the Lake Superior syncline in northwestern Wisconsin. It is similar to the gradient mapped over the known Douglas fault and suggests a considerable southwestward extension of the Lake Owen thrust. The center of the syncline has been thrust upward between the two faults as a horst.

(5) The gravity method can be applied to provide an estimate of thickness of lava and sandstone accumulations, attitude of contacts, and dip of faults. The method is therefore a valuable adjunct to geological mapping and interpretation in this region where only a small per cent of the area is exposed in outcrop.

References Cited

Aldrich, H. R., 1929, The geology of the Gogebic iron range of Wisconsin: Wis. Geol. Nat. Hist. Survey Bull. 71, 279 p.

Athy, L. F., 1930, Density, porosity, and compaction of sedimentary rocks: Am. Assoc. Petroleum Geologists Bull., v. 14, p. 1–24.

Birch, F., Schairer, J. F., and Spicer, H. C., 1942, Handbook of physical constants: Geol. Soc. America Special Paper 36, 325 p.

Bowie, W., 1917, Investigations of gravity and isostasy: U. S. Coast Geod. Survey Special Pub. 40, 196 p.

Buckley, E. R., 1898, Building and ornamental stones of Wisconsin: Wis. Geol. Nat. History Survey Bull. 4, 544 p.

Chamberlin, R. T., 1935, Geologic analysis of the gravity anomalies for the Black Hills-Bighorn-Beartooth region: Geol. Soc. America Bull., v. 46, p. 393–408.

Duerksen, J. A., 1949, Pendulum gravity data in the United States: U. S. Coast Geod. Survey Special Pub. 244, 218 p.

Evans, P., and Crompton, W., 1946, Geological factors in gravity interpretation illustrated by evidence from India and Burma: Geol. Soc. London Quart. Jour., v. 102, p. 211–249.

Fuller, M. L., 1906, Total amount of free water in the earth's crust: U. S. Geol. Survey Water-Supply Paper 160, p. 59–72.

Garland, G. D., 1950, Interpretations of gravimetric and magnetic anomalies on traverses in the Canadian Shield in northern Ontario: Pub. Dominion Observatory, v. 11, 57 p.

Glennie, E. A., 1932, Gravity anomalies and the structure of the earth's crust: Survey of India Prof. Paper 27.

Grout, F. F., and Schwartz, G. M., 1939, The geology of the anorthosites of the Minnesota coast of Lake Superior: Minn. Geol. Survey Bull. 28, 119 p.

Grout, F. F., *et al.*, 1932, Geologic map of Minnesota: Scale 1:500,000, Minn. Geol. Survey.

Gunn, R., 1949, Isostasy—extended: Jour. Geology, v. 57, p. 263–279.

Hammer, S., 1939, Terrain corrections for gravimeter stations: Geophysics, v. 4, p. 184–194.

Heiskanen, W., 1953, Geophysical applications of gravity anomalies: Am. Geophys. Union Trans., v. 34, p. 11–15.

Hoskins, L. M., 1896, Flow and fracture of rocks as related to structure: U. S. Geol. Survey 16th Ann. Rept., pt. 1, p. 845–875.

Hotchkiss, W. O., 1915, Mineral land classification: Wis. Geol. Nat. History Survey Bull. 44, 378 p.

—— 1923, The Lake Superior geosyncline: Geol. Soc. America Bull., v. 34, p. 669–678.

Hotchkiss, W. O., and Bean, E. F., 1929, Mineral lands of part of northern Wisconsin: Wis. Geol. Nat. History Survey Bull. 46, 212 p.

Hubbert, M. K., 1948, Gravitational terrain effects of two dimensional topographic features: Geophysics, v. 13, p. 226–254.

Irving, R. D., 1880, General geology of the Lake Superior region: Geology Wis., v. 3, pt. 1, Madison, Wis., p. 1–24.

—— 1883, The copper bearing rocks of Lake Superior: U. S. Geol. Survey Mon. 5, 464 p.

Julien, A. A., 1880, Microscopic examination of eleven rocks from Ashland County, Wisconsin: Geology Wis., v. 3, app. B, Madison, Wis., p. 224–238.

King, P. B., *et al.*, 1944, Tectonic map of United States: Am. Assoc. Petroleum Geologists, Scale 1:2,500,000.

Leith, C. K., Lund, R. J., and Leith, A., 1935, Pre-Cambrian rocks of the Lake Superior region: U. S. Geol. Survey Prof. Paper 184, 34 p.

Lyons, P. L., 1950, A gravity map of the United States: Tulsa Geol. Soc. Digest, v. 18, p. 33–43.

Martin, L., 1911, Physical geography of the Lake Superior region, p. 85–117, *in* Van Hise, C. R., and Leith, C. K., The geology of the Lake Superior region: U. S. Geol. Survey Mon. 52, 641 p.

Maxwell, J. C., and Verrall, P., 1954, Low porosity may limit oil in deep sands: World Oil, v. 138, no. 5, p. 106–113; no. 6, p. 102–104.

Melcher, A. F., 1921, Determination of pore space of oil and gas sands: Am. Inst. Min. Metall. Eng. Trans., v. 65, p. 469–489.

Nettleton, L. L., 1934, Fluid mechanics, of salt domes: Am. Assoc. Petroleum Geologists Bull., v. 18, p. 1175–1204.

Peters, J. W., and Dugan, A. F., 1948, Gravity and magnetic investigations at the Grand Saline Salt Dome, Van Zandt Co., Texas, p. 105–120, *in* Nettleton, L. L. (*Editor*), Geophysical case histories: Soc. Exploration Geophysicists, 671 p.

Pumpelly, R., 1880, Lithology of the Keweenawan system: Geology Wis., v. 3, pt. 2, Madison, Wis., p. 27–49.

Schwartz, G. M., 1944, Tracing the Duluth gabbro contact with a magnetometer: Econ. Geology, v. 39, p. 224–233.

—— 1949, The geology of the Duluth metropolitan area: Minn. Geol. Survey Bull. 33, 136 p.

Thwaites, F. T., 1912, Sandstones of the Wisconsin coast of Lake Superior: Wis. Geol. Nat. History Survey Bull. 25, 117 p.

——1935, Kansas Geol. Soc. Guidebook, 9th Ann. Field Conf., p. 221–228.

Van Hise, C. R., and Leith, C. K., 1911, The geology of the Lake Superior region: U. S. Geol. Survey Mon. 52, 641 p.

Van Orstrand, C. E., 1951, Observed temperatures in the earth's crust, p. 107–149, *in* B. Gutenberg (*Editor*), Internal constitution of the earth: N. Y., Dover Pubs., 439 p.

Vening Meinesz, F. A., Umbgrove, J. H. F., and Kuenen, Ph. H., 1934, Gravity expeditions at sea, v. 2: Netherlands Geodetic Commission, Delft, 208 p.

Welch, G. I., 1941, Geophysical study of the Douglas fault, Pine County, Minnesota: Jour. Geology, v. 49, p. 408–413.

White, D., 1924, Gravity observations from the standpoint of the local geology: Geol. Soc. America Bull., v. 35, p. 207–277.

Wisconsin Geological and Natural History Survey, 1939, Geologic map of Wisconsin, revised by E. F. Bean, Scale 1:1,000,000.

Woollard, G. P., 1938, The effect of geologic corrections on gravity anomalies: Am. Geophys. Union Trans., v. 19, p. 85–90.

—— 1943, Transcontinental gravitational and magnetic profile of North America and its relation to geologic structure: Geol. Soc. America Bull., v. 54, p. 747–790.

——1951, Annual report of the special committee on the geophysical and geological study of continents, 1950–1951: Am. Geophys. Union Trans., v. 32, p. 634–647.

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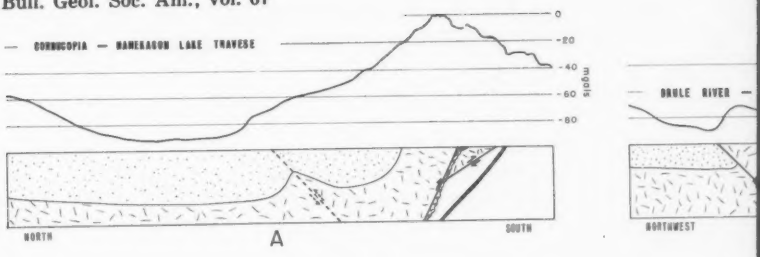
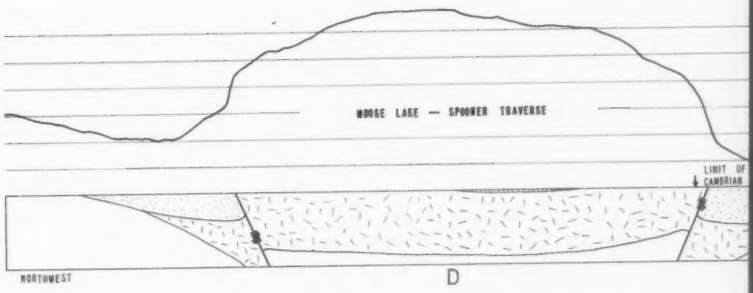
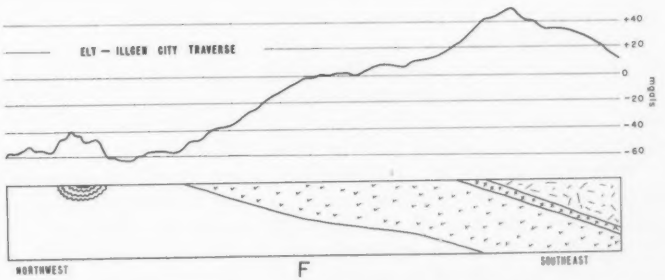
MANUSCRIPT RECEIVED BY THE SECRETARY OF THE SOCIETY, JUNE 21, 1955







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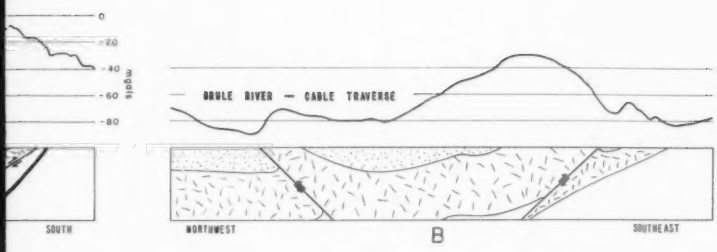

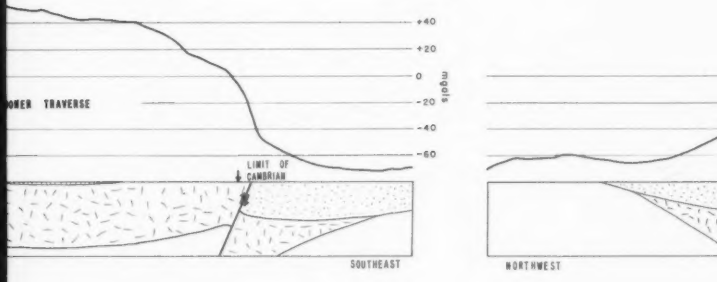

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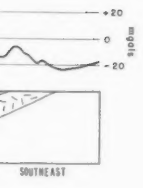

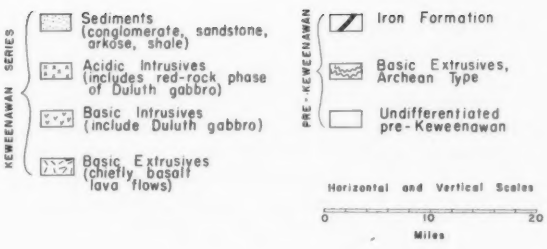

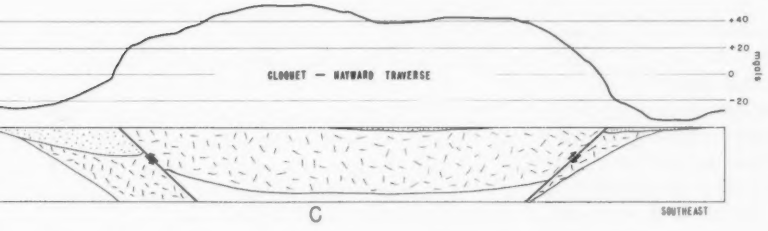

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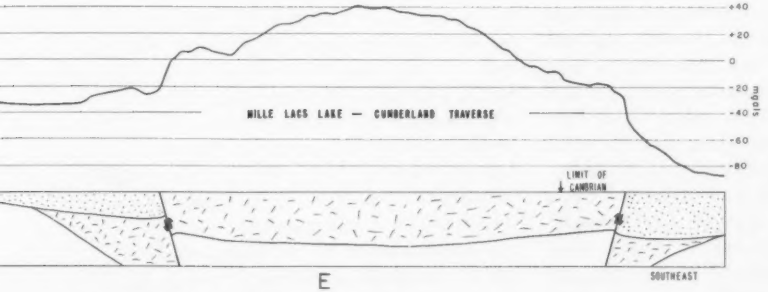

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KEWEENAWAN GEOLOGY

Short Notes

BULLETIN OF THE GEOLOGICAL SOCIETY OF AMERICA
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USE OF INDICATORS IN THE DETERMINATION OF ICE-MOVEMENT DIRECTIONS IN ALBERTA, CANADA: A DISCUSSION[1]

BY A. MACS. STALKER AND B. G. CRAIG

Introduction

Gravenor and Bayrock (1955, p. 1325) draw attention to the importance of "indicators" in determining general direction of ice movement. Such indicators are good evidence of general direction of movement of a single or of successive glaciers moving in the same general direction provided the possible sources of any single indicator can be limited to one district. However, the writers believe that the picture of ice movement on the western prairies of Canada, as obtained by Gravenor and Bayrock through use of indicators, is much oversimplified.

Gravenor and Bayrock do not state whether they refer to combined movement of all the Pleistocene glaciers in Alberta or only to movement of the last major glacier. If they refer to the resultant of the movement of all the glaciers the lines of general movement shown in their Figure 1 (p. 1326) are likely to be close to the true direction. However, they mention (p. 1325) glacial flutings in the Red Deer district of Alberta as corroborating evidence for the direction of ice movement determined from the indicators. These ice-flow markings were formed by the last major glacier in the area, and it is therefore assumed that they refer only to movement of this ice sheet. Other such markings elsewhere in Alberta and in Saskatchewan raise doubt as to whether the direction they obtain represents the true movement of the last glacier. In addition most other glacial erratics mentioned elsewhere in the literature indicate directions of ice movement that conflict with those obtained by Gravenor and Bayrock. Some, though by no means all, of the known ice-flow markings and indicator erratics on the Canadian plains are described here. The numbers in parentheses correspond to those on Plate 1.

Evidence of Ice-Flow Markings

The prominent flutings in the Red Deer district (1) are in the pre-glacial valley of the Red Deer River, much of which has been abandoned by the modern stream. This former valley is about 15 miles wide and several hundred feet deep below the high land that marks its eastern edge and locally its western edge. This valley changed the direction of local ice movement, and the flutings mentioned by Gravenor and Bayrock represent to some extent the strike of this valley and not necessarily the general direction of glacier flow. This strong control by local topography of movement of the last glacier is characteristic in much of central and southwestern Alberta, and only examination of ice-flow markings in an extensive area gives a correct picture of overall general movement. The strike of these markings can change as much as 20° in 4 miles and 100° in 50 miles. Several prominent occurrences of such markings are mentioned here.

Some ice-flow markings (2) west of Ponoka, and a few miles north of the flutings to which Gravenor and Bayrock refer, indicate movement directly to the south. Other such markings (3) east of Ponoka and 25 to 40 miles northeast of the Red Deer flutings indicate ice movement of S. 22°–42° E. Similar markings (4) between 10 and 40 miles south of those in the Red Deer district indicate ice

[1] Published by permission of the Deputy Minister, Department of Mines and Technical Surveys, Ottawa.

movement of S. 8°–25° E. Farther southward and southeastward the easterly component increases, and the markings (5) indicate local ice movement to as much as S. 60° E. Various southerly movements are indicated by ice-flow markings (6) 40 or 50 miles east of those in the Red Deer district. The writers have studied such markings in much of southwestern and central Alberta, in addition to stone orientation in till, and have traced a major boulder train in an attempt to determine general direction of ice movement. They believe that in central Alberta the last major glacier moved southward to southeastward and in southwestern Alberta somewhat parallel to the Rocky Mountain front. Movements that apparently conflict with this overall direction of movement generally are of limited extent and controlled by local topography.

Ice-flow markings (7) are present in the North Battleford district of Saskatchewan, near some of Gravenor's and Bayrock's easternmost lines of general movement. Many of the most prominent of these indicate a general southeastward movement, though the direction indicated by some of the others is much different. Kupsch (1955, p. 330, 331, 336) describes drumlins (8) in the Dollard district of Saskatchewan that indicate ice movement of about S. 30° W. This district is just southeast of the Cypress Hills, which probably had strong influence on the ice flow, and this southwestward movement may have been only local. Other ice-flow markings in western Manitoba and southern and central Saskatchewan present a somewhat confused picture of ice movements, but the strongest trend is to the southeast (J. A. Elson, unpublished map).

Evidence of Indicators

The evidence of ice-movement direction given by indicators, including those mentioned by Gravenor and Bayrock and others described elsehwere in the literature, is confused. More than one source area is possible for many of the erratics (*i.e.*, some sandstones and oil-impregnated dolomites), and the outcrop area of others (*e.g.*, Presqu'ile dolomite) is large or not yet defined.

Rutherford (1941, p. 119), in discussing erratics in central Alberta, states:

> "The sources of only two rock types in the moraines of central Alberta are known, if we assume a limited distribution of those rocks in place. These are from the bituminous sands at McMurray and the Presqu'ile dolomite at the western end of Great Slave Lake. The dolomite, possessing pores filled with residual hydrocarbons, is fairly common in the moraines in the Edmonton district. The bituminous sand boulders and derived material occur in the western moraines between Saskatchewan and Athabaska Rivers. These rock types suggest a northerly source for the material of the earlier moraines".

If they came from the western end of Great Slave Lake, as suggested by Rutherford, these dolomite boulders (9) indicate a general ice movement of a few degrees east of south. If they came from Pine Point, the source Gravenor and Bayrock suggest for their dolomite indicator, they indicate movement nearly to the south. The writers have noted similar oil-impregnated dolomite (10) a few miles east of Olds, Alberta. If this came either from Pine Point or the western end of Great Slave Lake the movement indicated would also be southward or slightly east of southward. Kupsch (1954) notes bituminous dolomite boulders (11) in the Peter Pond Lake area of Saskatchewan. If these came from the Presqu'ile dolomites they indicate general ice movement to the southeast or south-southeast. Kupsch mentions (p. 20) that lithologically similar rocks are found *in situ* in the Devonian strata of the Clearwater valley at several places—*e.g.*, Pas Rapid—and that "They have also been collected farther north where the Devonian is not overlain by Cretaceous sediments." These sources would indicate southeastward or south-southeastward movement of the ice.

Gravenor and Bayrock (p. 1327) note the large area of outcrop of Athabasca sandstone and mention that lines from the indicators to various parts of this outcrop can give "slightly different directions of ice movement." Several of the easternmost indicators shown could have moved directly southward. Kupsch (1954, p. 20, 33, 34) also reports boulders (11) similar in type to the Athabasca sandstone, and practically certain to have come from this formation, in the Peter Pond Lake area. These indicate ice movement to the south, though much variation is possible as the Athabasca sandstone is exposed in a large area. Other erratic boulders of type similar to the Atha-

basca sandstone are common in much of the prairies. J. A. Elson (Oral communication) has noted many such sandstone fragments (12) in southwestern Manitoba and southeastern Saskatchewan, between 49° and 50° N. Lat. and 101° and 102° W. Long. These, if Athabasca sandstone, indicate ice movement to the southeast or south-southeast.

The two boulders (9) described by Rutherford (1928) and mentioned by Gravenor and Bayrock undoubtedly came from the bituminous sand of the McMurray district, and, as Rutherford states (p. 561, 562), these boulders represent movement of S. 15°–35° W. in one case and S. 20°–40° W. in the other. Kupsch (1954, p. 20, 21) reports blocks (11) of such bituminous sand in the Peter Pond Lake area, and he infers that they came from the McMurray tar sands. No other derivation seems possible for these blocks, and they indicate ice movement to the southeast, in a direction greatly different from that travelled by the boulders of Rutherford.

Sandstone is a common rock in the Rocky Mountains west of the plains, and some of this sandstone is much like the Athabasca sandstone. Local facies of this sandstone similar to some of the Athabasca sandstone are undoubtedly present, and the use of sandstone erratics as indicators in the western part of the prairies, particularly if they are small, is risky. The most that can be inferred about the origin of many sandstone erratics on the Alberta prairies is that they could have come from the Athabasca sandstone. Gravenor's and Bayrock's valuable study of the sandstone erratics greatly increases this likelihood. The possibility remains, however, that some of them, and particularly those in the western part of the prairies, had a different source, and, as Gravenor and Bayrock state, the identification is not positive.

The use in Alberta of glacial erratics as indicators of direction of ice movement thus results in a large range of possible movements. In general the indicated direction in central Alberta is southward; farther east, in central Saskatchewan, the indicated direction is southeastward. These directions agree with those derived from study of ice-flow markings.

Value of Indicators

The discrepancies in the directions of ice movement shown by the indicators mentioned, by those mentioned by Gravenor and Bayrock, and by ice-flow markings raise the question of the value of indicators in a region glaciated by successive ice sheets that moved in different directions. This evidently happened in Alberta, where the successive glaciers may have varied in direction of movement by as much as 60°. Indicators are very valuable in an area that was glaciated only once, or by successive ice sheets that advanced in the same direction. In Alberta, however, it is doubtful whether they give any indication, except a very rough one, of general direction of combined movement of all the glaciers. An erratic in Alberta that was moved a long distance by the first glacier, but very little by succeeding ones, would certainly indicate a direction of ice movement different from that indicated by an erratic that, coming from the same location, was moved only by a later glacier moving in a direction different from that of the first glacier. To be a good indicator of the resultant of the general movements of all the glaciers in Alberta an erratic must have been moved by each glacier a distance in direct and equal ratio to the total movement of each glacier. Thus if the first glacier carried a block half the distance of the glacier's total movement, a second glacier moving in a different direction would also have to move this block half the distance of its movement in order that the block give a true resultant movement of the two ice sheets. Few of the erratics in Alberta have been moved in this manner, and few of them give a correct indication of the combined general movement of all the glaciers. However, if a single glacier moved an erratic the whole way to its present location, that erratic is a true indicator of direction of movement of that glacier. This is generally impossible to prove.

Forthcoming maps of surficial deposits in various areas will show many ice-flow markings, and knowledge of ice movement on the plains will increase rapidly. It may be best to leave the problem of ice movement until these maps are published and other information is available, and the writers think best not to answer in detail here the specific points raised in

Gravenor's and Bayrock's reply. The writers had carefully avoided mention of the moraines on the prairies, features they have studied intensively, because these moraines were not mentioned in the original paper. These moraines are a complex problem, and the writers expect to discuss certain of their features in a future paper, and some description of them is present in forthcoming publications. The writers also confined themselves to discussing those types of indicators mentioned by Gravenor and Bayrock. They felt that these alone were sufficient to show the conflicting evidence. The writers conformed with the outcrop area of the Athabasca sandstone shown on Gravenor's and Bayrock's map, rather than use the outcrop area shown on the new Geological Map of Canada, as it made little difference to the argument.

The writers originally started mapping on the western plains with the assumption, gained from much of the earlier literature, that the glacier had moved southwestward. After several years they realized this assumption might be wrong, and then only after careful study of many aerial photographs and features that could not be explained by southwestward movement. They finally came to the view expressed by one of the greatest early students of the western plains, J. B. Tyrrell, who, in discussing ice movement in the great Winnipeg valley and upper Assiniboine valley in Manitoba, stated (1890, p. 401) "In both these cases the direction of flow was southward or southeastward in the direction of the trend of the valleys, and parallel to the main axis of the Rocky Mountains. This direction was in all probability sustained by the glacier all the way across the Canadian Plains."

Conclusions

The direction reported by Gravenor and Bayrock for general ice movement is unlikely to be that of the last major glacier. It may be near the resultant movement of all the successive glaciers in Alberta, though this is not necessarily proved by the indicators used. The granite, gneiss, and schist erratics from the Canadian Shield, which are generally used as a criterion of Laurentide glaciation on the western prairies, certainly would not be present in southwestern Alberta unless the movement of one or more of the glaciers had a westerly component. If these erratics were brought from the parts of the Canadian Shield closest to their present locations, their general movement was between S. 20°W. and S. 60° W.

It is hoped that Gravenor and Bayrock will continue their study of the heavy minerals of the various erratics and their possible source rocks. Such study is certain to yield valuable information about the Pleistocene history of Alberta. Possibly they will extend their study to the crystalline erratics so common on the prairies, which definitely came from the north or northeast. The best indicators would be those from areas that apparently were glaciated only once. By such means they might determine the general directions of movement of each of the three or so successive major glaciers, and so discover which glacier had the greatest westward component to its movement. The question of which glacier brought the crystalline boulders to southwestern Alberta is of great importance in the unravelling of the glacial history of Alberta. It apparently was not the last major glacier.

References Cited

Gravenor, C. P., and Bayrock, L. A., 1955, Use of indicators in the determination of ice-movement directions in Alberta: Geol. Soc. America Bull., v. 66, p. 1325–1327.

Kupsch, W. O., 1954, Bituminous sands in till of the Peter Pond Lake area, Saskatchewan: Saskatchewan Geol. Survey Rept. 12, 35 p.

—— 1955, Drumlins with jointed boulders near Dollard, Saskatchewan: Geol. Soc. America Bull., v. 66, p. 327–338

Rutherford, R. L., 1928, Two interesting boulders in the glacial deposits of Alberta: Jour. Geology, v. 36, p. 558–563

—— 1941, Some aspects of glaciation in central and southwestern Alberta: Royal Soc. Canada Trans., v. 35, sec. 4, p. 115–124

Tyrrell, J. B., 1890, Post-Tertiary deposits of Manitoba and the adjoining territories of northwestern Canada: Geol. Soc. America Bull., v. 1, p. 395–410

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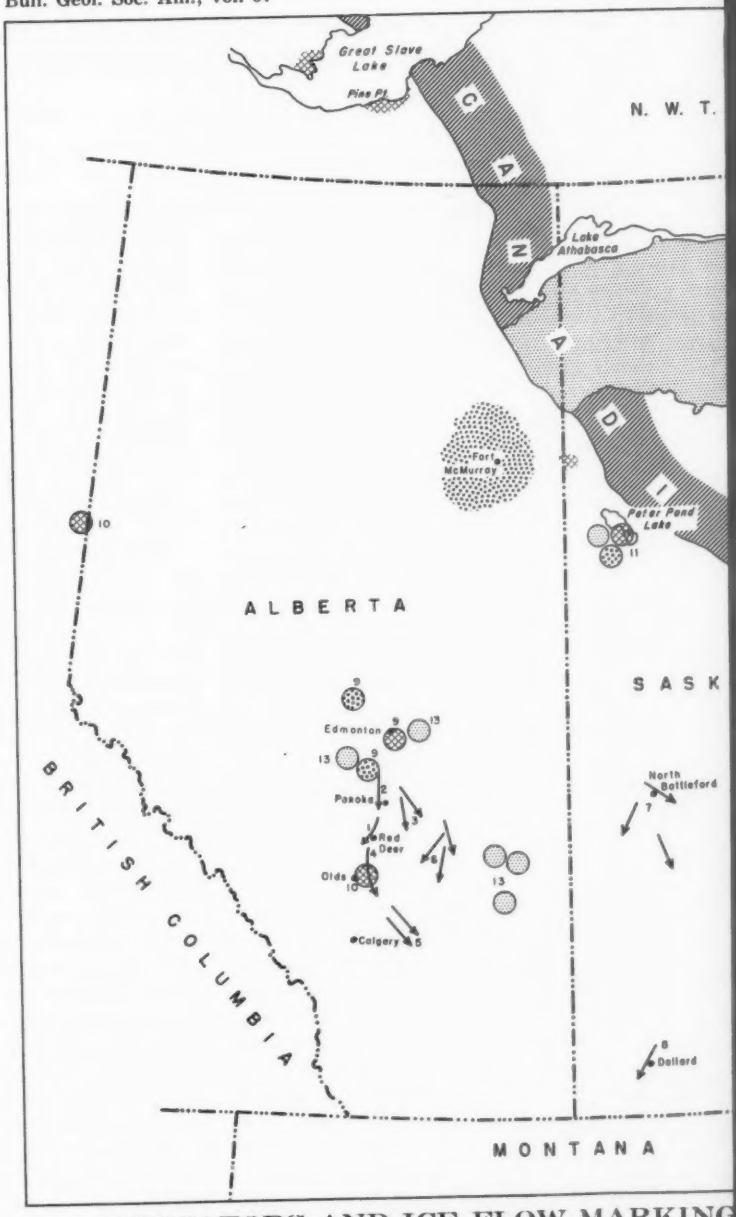
MANUSCRIPT RECEIVED BY THE SECRETARY OF THE SOCIETY, JANUARY 30, 1956



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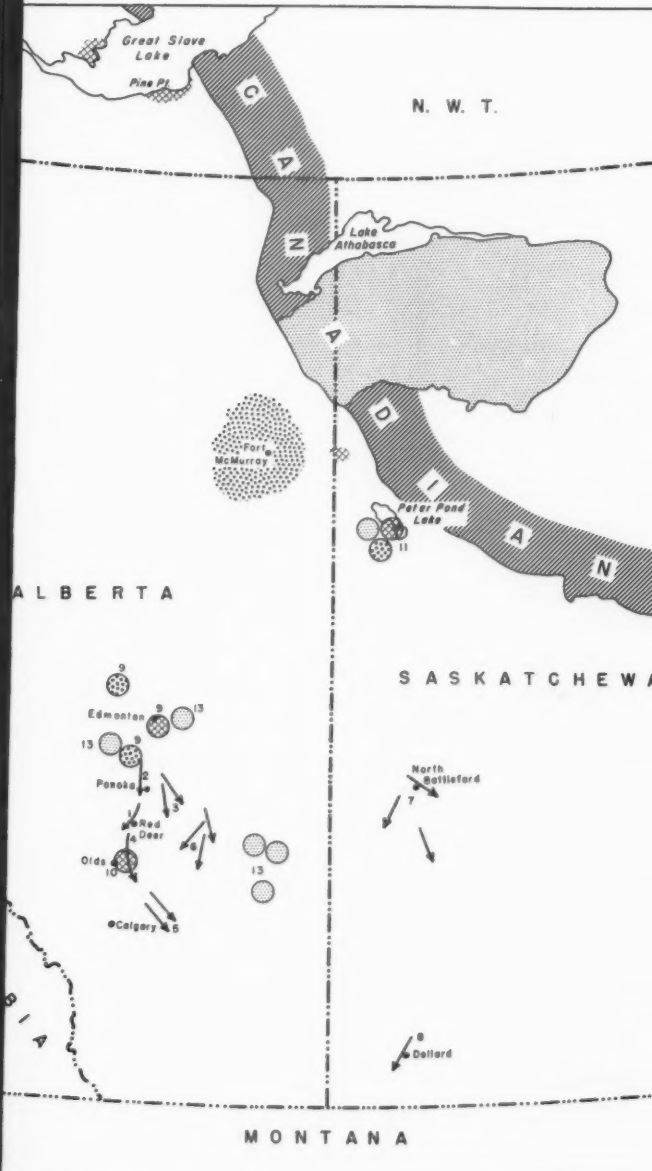

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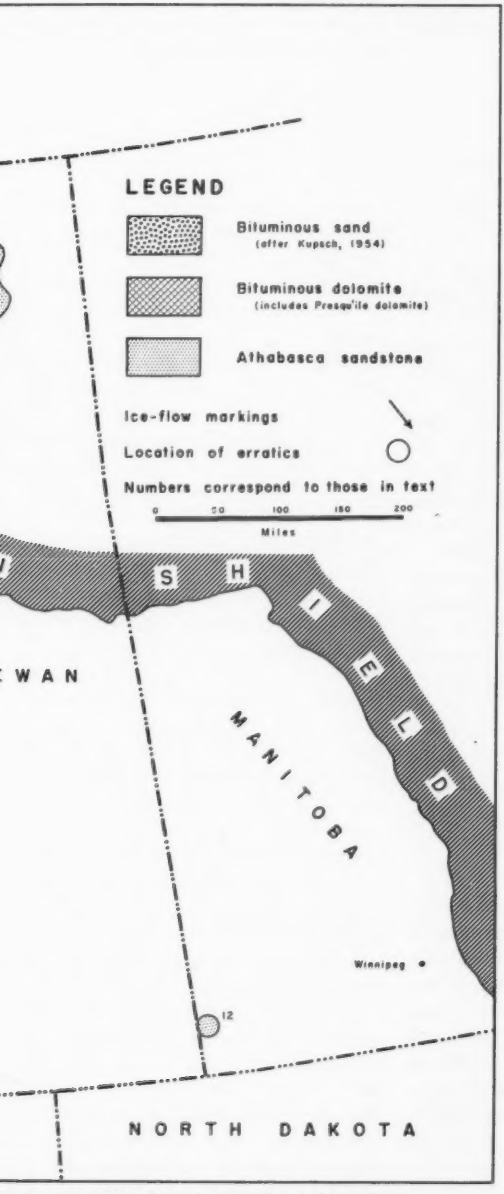

HE CANADIAN PRAIRIES

BULLETIN OF THE GEOLOGICAL SOCIETY OF AMERICA
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USE OF INDICATORS IN THE DETERMINATION OF ICE-MOVEMENT DIRECTIONS IN ALBERTA, CANADA: A REPLY

BY C. P. GRAVENOR AND L. A. BAYROCK

Introduction

In a discussion of ice-movement directions over a given region it is necessary to distinguish between local and regional directions. As Stalker and Craig (1956) point out, it is possible to interpret local ice directions in terms of general ice direction only when a great number of local directions are known. This point is well illustrated in southeastern Wisconsin where the configuration of the Johnstown moraine and Kettle interlobate moraine clearly shows that the Green Bay ice lobe moved southwesterly (Alden, 1918, p. 208). Drumlins, eskers, and striae on the west side of the Green Bay lobe show southwesterly, westerly, and even northwesterly directions; on the east side of the lobe they show southeasterly directions and on the southern part of the lobe they show southeasterly, southerly, and southwesterly directions. The appearance of several ice lobes in a given region enormously complicates the interpretation of local ice directions (Alden 1918, p. 194, Pl. 20). Thus two sets of local directions only 10 miles apart may belong to two different ice lobes and cannot be related to a general ice direction.

In southern Alberta and northern Montana the configuration of the Cary and Tazewell moraines (Fig. 1) shows the general ice-movement direction, and local directions can be interpreted in light of regional direction. Unfortunately, in central and northern Alberta the meager amount of information does not allow the interpretation of local glacial directions in terms of lobation. The writers are well aware of the difficulties inherent in the indicator method of determination of ice-movement direction; they are also aware of the fact that unless local ice directions can be fitted into a regional pattern they may give a very misleading account of general ice-movement directions. Until such time as the regional setting is available in all of central and northern Alberta, we suggest that indicators serve a useful purpose.

To the reader unacquainted with the publications on glacial geology of Alberta, Stalker and Craig's discussion (1956) might leave the impression that the writers were extremely biased in their approach and ignored that material which does not agree with their thesis. This is not the case. The writers did not mention certain occurrences of erratics such as the bituminous sands reported from the Peter Pond Lake area of Saskatchewan (Kupsch, 1954) because there is some doubt that these occurrences represent general ice movement in a southwesterly direction. Most publications on the glacial geology of Alberta admit to a general southwesterly ice-movement direction and in a Short Note the writers did not think it necessary either to review the literature or to submit controversial evidence. It now seems necessary, however, to clarify the situation by giving a brief summary of previous work on Alberta glacial geology. For the sake of brevity, the writers concentrate on articles which cover those areas in Alberta under discussion.

Previous Work

Stalker and Craig (1956) indicate that they believe the last major glacier in central Alberta moved in a southerly to southeasterly direction and that in southwestern Alberta the movement somewhat parallels the Rocky Mountain front—that is, presumably, in a southeasterly direction. They also indicate that in their opinion the crystalline boulders found in southwestern Alberta were not taken there by the last major glacier.

Horberg (1952) in his thorough study of the Pleistocene geology of the Lethbridge region of southern Alberta states (p. 307): "The fact that the Keewatin ice sheet had to advance upslope into the headwaters area of the Sas-

katchewan drainage basin is a fundamental consideration in interpreting both the earlier and the late Pleistocene history of the region." He also (p. 316) discusses the relationship of the upper or surface till in the Lethbridge area,

In a paper on the Pleistocene deposits of the Waterton area of southwestern Alberta, Horberg (1954, p. 1115) makes the following statement: "The areal relations of the continental moraines within the region show that

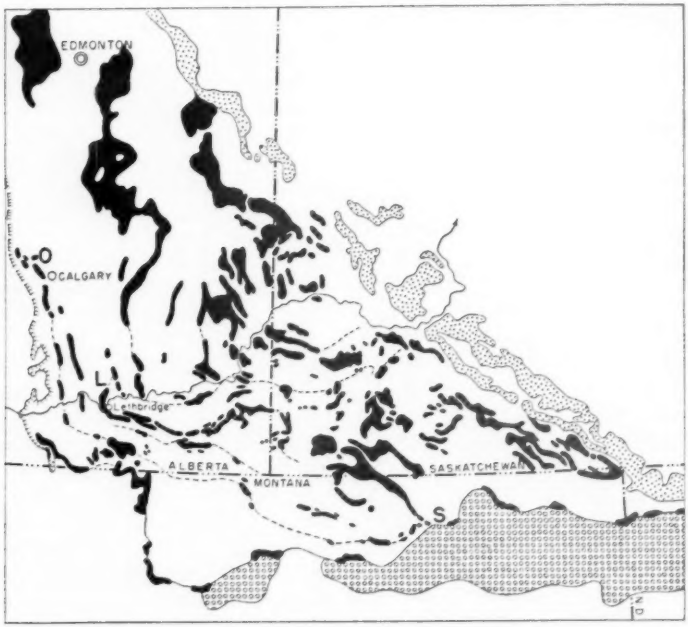

FIGURE 1.—MAP SHOWING GLACIAL DEPOSITS IN CENTRAL AND SOUTHERN ALBERTA AND ADJOINING REGIONS

Iowan (?) drift shown by circle pattern, Tazewell-Cary moraines in black, and Mankato Altamont moraine by dotted pattern. O, Okotoks moraine; L, Lethbridge moraine; S, Moraine overlap near Saco, Montana (Taken from Horberg, 1952)

"Correlation of the upper till depends largely upon one's evaluation of the time interval between deposition of the till and the underlying Lenzie silt. During this interval the ice retreated at least fifty miles east or thirty-five miles northeast of the area and then re-advanced to the position of the Lethbridge moraine." Another interesting observation made by Horberg (1952, p. 317) is that the upper till in the Lethbridge region contains more crystalline material than the lower tills. These statements and the last observation regarding crystalline material indicate that Horberg considered the general ice-movement directions in this part of Alberta to be westerly to southwesterly.

they were deposited by a major ice lobe that protruded westward from the position of the Lethbridge moraine into the broad, regional lowland between the Porcupine Hills on the north and the Rocky Mountain foothills and Milk River Hills to the south."

About this same general area Johnston and Wickenden (1931, p. 30) state: "In southern Alberta there are two drift sheets of the Laurentide or northeastern ice-sheet and these are separated in places by lignite-bearing interglacial deposits. The earlier ice-sheet appears to have advanced farther southwest and west than the later one and to have extended well into the foothill region."

In the foothills area north of Waterton,

Douglas (1950, p. 6–7) indicates that the Porcupine Hills played a major role in deflecting the ice of this region, and major and minor ice lobes moved north, south, and west.

Still farther north in the Jumping Pound area west of Calgary, Nichols (1931, p. 55) reports, "It would appear that the earlier Laurentide ice-sheet advanced from the northeast and reached a point at least thirty-five miles northwest of Calgary and to within 30 miles of the base of the mountains, passing some distance beyond the escarpment of the upper plateau and over an area which, previously, had been occupied by a Cordilleran ice-sheet. The later advance of the glaciers from the east did not reach as far west as the earlier one, nor to as high an altitude, halting about the western edge of the lower plateau."

If we accept the ideas of Horberg, Johnstone and Wickenden, Douglas, and Nichols, then the major ice-movement direction in southern and southwestern Alberta was westerly to southwesterly, and variations from this general movement are the result of local tongues of ice. If we accept the age relationship of the moraines in this region as presented by Horberg (1954), then the crystallines found in south western Alberta were brought there by the last major ice to occupy southern Alberta.

In a discussion of some glacial features of central Alberta, Warren (1954, p. 84) states, "The silt till is the youngest till in central Alberta and it appears to be quite restricted in area. The glacier that deposited it advanced from the northeast, passing to the west of Cooking Lake Hills." Of the Duffield moraine west of Edmonton, Warren (1954, p. 83–84) states, "It is evident that the glacier that deposited the silt till and was responsible for the Duffield moraine crossed the area from the north or northeast. This is shown by the strike of the drumlins in the vicinity of Edmonton and by the direction of elongation of the elongated hills in the morainal area."

In the St. Ann area west of Edmonton, Collins and Swan (1955) state, "The till deposits studied are all of Keewatin origin and represent debris accumulating at or near the maximum westerly advance of glacial ice sheets."

The moraines of central Alberta are very broad and do not show such a clear picture of lobation as those of southern Alberta. However, the Altamont and Viking moraines of east-central Alberta trend southeasterly and might be interpreted as having been deposited from a southwesterly flowing ice. Bretz (1943, p. 34) considers the four Keewatin moraines in central Alberta in the general atitude of Edmonton to be younger froml west to east and of the Duffield moraine states (p. 36), "The Keewatin ice sheet doubtless lay immediately to the east when this moraine was deposited, but the cause of its western front is not so clear." Bretz clearly indicates that he believes the moraines of central Alberta were deposited by westerly to southwesterly flowing ice; certainly it does not seem likely that the moraines could be younger from west to east if the last ice advanced from the northeast.

The writers do not attach too much significance to the lineation of the moraines of central Alberta, since the conditions of deposition are not as yet well understood. Some of these moraines, especially the Duffield moraine, may be interlobate or the result of stagnation of debris-choked ice. These remain only speculations until the regional picture is clearer. It is hoped that the work of Stalker and Craig will help to clear up this most complex problem.

In mentioning the drumlins of the Dollard district of Saskatchewan (Kupsch, 1955) which give southwesterly ice-movement directions, Stalker and Craig (1956) state, "This district is just southeast of the Cypress Hills, which probably had a strong influence in the ice-flow, and this southwestward movement may have been only local."

Although we doubt if the ice-movement direction of the Cypress Hills area has much to do with directions in central and northern Alberta, it is nevertheless rather interesting to note statements made by other geologists about this area.

In a study of the geology of southern Alberta and Saskatchewan, Williams and Dyer (1930, p. 110) state: "Moraines, kettles, boulders and a general glaciated topography characterize the upland south of Ravenscrag and East End, indicating that an extensive ice-lobe diverted south by the elevation of the Cypress Hills, covered this region extending in a southwest

direction." In Johnston and Wickenden's report (1931, p. 39) on the glacial geology of southern Saskatchewan and southern Alberta, they state, "As the ice-sheet moved towards the southwest and must have had some surface slope in the direction of movement, its surface probably stood above the tops of the Cypress Hills, though it did not over-ride them or deposit any drift on their upper parts." These two statements suggest that the southwesterly trending drumlins in the Cypress Hills area represent a regional ice-flow pattern. Other evidence from south-central Saskatchewan and north-central Montana also suggests regional southerly and southwesterly and local southeasterly movements (Knechtel, 1942).

Indicators

Athabasca sandstone.—Stalker and Craig (1956) suggest that sandstone erratics found in southwestern Manitoba were derived from the Athabasca sandstone outcrops on the south side of Lake Athabasca. Although no formal mineralogical description of these erratics is given, if it is assumed that they are Athabasca sandstone indicators, then lines drawn from southwestern Manitoba to the outcrop area of the Athabasca sandstone give directions of ice movement which are in conflict with most of the known directions as shown on the Glacial Map of North America (Flint, 1945).

Stalker and Craig (1956) state that some of the easternmost Athabasca sandstone indicators in Alberta could have been derived from the western edge of the Athabasca sandstone outcrop and consequently could give north-south ice-movement directions. Sproule (1951) shows that the western edge of the Athabasca sandstone outcrop area lies farther east than is shown on the Geological Map of Canada. Recent mapping by the Research Council of Alberta (in preparation) supports Sproule and indicates a more easterly border to the Athabasca sandstone. Hence a general southwestern direction is indicated even for the easternmost Athabasca sandstone indicators in Alberta.

As Stalker and Craig state, other sandstones in the plains—especially some of the Tertiary quartzitic boulders—resemble Athabasca sandstone. In the Short Note the writers aimed to describe a technique which would aid in separating these sandstones. They do not suggest that the method is foolproof, but it should be considered as contributory evidence.

Indicators in the Peter Pond Lake area.—Stalker and Craig (1956) state that the only possible point of origin for the bituminous sands found in the Peter Pond Lake area is in the tar sand area of Alberta. There are at least two other possible explanations for this occurrence: (1) oil-saturated Cretaceous sands may exist, or may have existed, prior to glaciation (Sproule 1951, p. 12), between the Peter Pond Lake area and the Precambrian Shield (Wickenden *in* Kupsch, 1954, p. 25), and (2) blocks of oil-saturated sands may have been rafted out of the Churchill River system and subsequently incorporated in the drift of the Peter Pond Lake (Sproule *in* Kupsch, 1954, p. 28). Although Kupsch (1954) argues against these possibilities, the lack of knowledge of the subsurface rocks of this area makes it rather difficult to state unequivocally that the tar sands of Alberta were the source rocks.

A regional southwest ice-movement direction for the Peter Pond Lake area and northeastern Alberta is strongly supported by ice directions reported from neighboring regions. Immediately north and northeast of the Peter Pond Lake area Sproule (1939) reports a large tract of drumlins and eskers which indicate a southwesterly ice-movement direction. Directions to the east of Peter Pond Lake shown on the Glacial Map of North America (Flint, 1945) also show southwesterly directions. The writers have examined air photographs of the area between the Clearwater and Firebag rivers in Alberta (northeast of the Peter Pond Lake area), and drumlins and groovings of this region and in the Birch Mountains in general agree with the directions reported by Sproule. This mass of southwesterly directions which form a semicircle about the Peter Pond Lake area strongly suggest that the general ice-movement direction in this region is southwesterly.

However, even if the tar sands were derived from the McMurray area of Alberta, this

should not alter the picture. As ice expands over an area, indicators are usually distributed in a fan shape—similar to the ice-movement directions displayed by an expanding ice lobe. As an ice lobe started to expand in a general southwesterly direction over the McMurray area, possibly movement would be southeasterly on the southern side of the lobe. Thus, the tar-sand erratics found at Peter Pond Lake could be the result of the early phases of a southwesterly expanding ice lobe in this region.

In addition to the two occurrences of bituminous sands found in the Edmonton district and reported by Rutherford (1928), Feniak (1944, p. 18) reports bituminous erratics at Mosside, Hondo, Vega, and southeast of Jarvie. Feniak suggests that these bituminous sand erratics had their source in the McMurray area 200 miles to the northeast.

Presqu'ile dolomite.—As Stalker and Craig (1956) suggest, the oil-impregnated and bituminous dolomites found in central Alberta may have had more than one source. For this reason, the writers did not use these dolomite erratics as indicators.

The oil-bearing dolomite found in the Pouce Coupe region of British Columbia, however, presents a different case. All the known possible source areas lie east and northeast of Pouce Coupe. Since regional westerly ice-movement directions across this part of Alberta seem rather unlikely, a northeast source (Pine Point on Great Slave Lake) is probable. Thus geographic position dictates whether or not a particular rock type may be used as an indicator.

Southeast of Great Slave Lake, ice-movement directions are westerly to southwesterly (Flint, 1945). In northeastern British Columbia, in the Beatton River area, Matthews (1956) shows ice-movement directions which trend southwest-northeast. South of this area and extending into the Peace River country of Alberta, Matthews shows south and slightly east of south directions. These directions might be interpreted to mean that the ice moved south westerly into northeastern British Columbia and, upon meeting the Cordilleran ice front, turned to the south.

Thus, it is possible that the Presqu'ile dolomite was picked up by the ice from the Pine Point area and moved westerly or southwesterly across northwestern Alberta in the region of Hay Lakes and then more southerly to the Pouce Coupe district. This is, of course, a rather long-range extrapolation of ice directions, and until more is known about directions in northwestern Alberta the writers have indicated the conventional straight line direction between the indicator found at Pouce Coupe and the probable source area at Pine Point.

Comments on glacial directions in Central Alberta.—As Stalker and Craig point out, topography probably played a strong hand in determining the directions of ice movement in that portion of Alberta covered in their discussion. The extreme parallelism and length (30–40 miles) of the flutings found in the Red Deer district, suggested to the writers that they were formed under very thick ice which would not have been seriously affected by the preglacial channel. The interpretation of these markings, as well as others in this general region, would seem to be, in part, subjective.

Although the writers welcome the new information provided by Stalker and Craig—and we have no objection to local southerly or southeasterly directions in parts of central Alberta—to extrapolate this information into other parts of central Alberta would be rather hazardous. Consider, for example, the markings shown east of Calgary (Stalker and Craig, 1956, No. 5) which trend in a strong southeasterly direction. If these directions are placed on a map showing the morainal configurations in southern Alberta, they might well be interpreted as northwesterly directions, developed on the northwestern side of the ice lobe which occupied southern Alberta. Similarly, if a major ice lobe moved across the McMurray area in a southwesterly direction and entered north-central Alberta, then the markings shown by Stalker and Craig (1956, Nos. 2, 3, 4) might represent either the movements of the southern edge of this lobe or a swinging of the lobe to a more southerly route. In either case the regional direction of the lobe passing over northeastern and north-central Alberta would be southwesterly.

The writers do not suggest that the above interpretations are the only ones and, lest it

appear that we are trying to fit all directions into a preconceived plan, they are prepared to accept regional southerly to southeasterly directions over all of central Alberta when a sufficient number of local directions suggest this to be the case. Until east-central, north-central, and northern Alberta have been adequately examined and the regional lobation worked out, the writers suggest that indicators serve a useful purpose in giving general ice directions.

References Cited

Alden, W. C., 1948, The Quaternary geology of southeastern Wisconsin with a chapter on the older rock formations: U. S. Geol. Survey Prof. Paper 106, p. 1–356

Bretz, J H., 1943, Keewatin end moraines in Alberta, Canada: Geol. Soc. America Bull., v. 54, p. 31–52

Collins, G. A., and Swan, A. G., 1955, Glacial geology of the St. Ann area: Research Council of Alberta Rept. No. 67

Douglas, R. J. W., 1950, Callum Creek, Langford Creek and Gap map-areas, Alberta: Geol. Survey Canada Mem. 255

Feniak, M., 1944, Athabasca-Barrhead map-area, Alberta: Geol. Survey Canada Paper 44-6

Flint, R. F., *et al.*, 1945, Glacial map of North America: Geol. Soc. America Special Paper 60

Horberg, Leland, 1952, Pleistocene drift sheets in the Lethbridge region, Alberta, Canada: Jour. Geology, v. 60, p. 303–330

—— 1954, Rocky Mountain and continental deposits in the Waterton region, Alberta, Canada: Geol. Soc. America Bull., v. 65, p. 1093–1150

Johnston, W. A., and Wickenden, R. T. D., 1931, Moraines and glacial lakes in southern Saskatchewan and southern Alberta, Canada: Royal Soc. Canada Trans., v. 25, sec. 4, p. 29–44

Knechtel, M. M., 1942, Snake Butte boulder train and related glacial phenomena, north-central Montana: Geol. Soc. America Bull., v. 53, p. 917–936

Kupsch, W. O., 1954, Bituminous sands in till of the Peter Pond Lake area, Saskatchewan: Saskatchewan Geol. Survey Rept. 12

—— 1955, Drumlin with jointed boulders near Dollard, Saskatchewan: Geol. Soc. America Bull., v. 66, p. 327–338

Matthews, W. H., 1956, Map 5, British Columbia Natural Resources Atlas: British Columbia Nat. Resources Conf. (in press)

Nichols, D. A., 1931, Terminal moraines of the Pleistocene ice-sheets in the Jumping Pound, Wildcat Hills area, Alberta, Canada: Roy. Soc. Canada Trans., v. 25, sec. 4, p. 49–59

Rutherford, R. L., 1928, Two interesting boulders in the glacial deposits of Alberta: Jour. Geology, v. 36, p. 558–563

Sproule, J. C., 1939, The Pleistocene geology of the Cree Lake region, Saskatchewan: Roy. Soc. Canada Trans., v. 33, sec. 4, p. 101–109

—— 1951, The McMurray formation in its relation to oil occurrence: Athabasca Oil Sands Conf., Edmonton, Proc., p. 6–29

Stalker, A. MacS., and Craig, B. G., 1956, Use of indicators in the determination of ice-movement direction in Alberta—A discussion: Geol. Soc. America Bull., v. 67, p. 1101–1104.

Warren, P. S., 1954, Some glacial features of central Alberta: Roy. Soc. Canada, Trans., v. 48, sec. 4, p. 75–85

Williams, M. Y., and Dyer, W. S., 1930, Geology of southern Alberta and southwestern Saskatchewan: Geol. Survey Canada Mem. 163

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APPARENT ABSENCE OF SOIL CREEP IN THE EAST GEORGIA PIEDMONT

BY ELDON J. PARIZEK AND JAMES F. WOODRUFF

Prior to 1938, when C. F. S. Sharpe published the classic study *Landslides and related phenomena*, mass wasting or gravitative transfer of material had not received adequate attention and analysis as a geomorphic process. Sharpe proposed a new and more detailed classification of mass wasting which has been widely adopted by most recent writers.

The primary basis for the classification is the "kind of movement"—*i.e.*, flowage or slippage—and its "relative rate" (Sharpe, 1938, p. 17). Further subdivisions are based upon "relative water or ice content" and "kind of material" involved. Included within the slow-flowage category are rock creep, talus creep, soil creep, rock-glacier creep, and solifluction. Sharpe believes that "in surveying the types of mass movements . . . it becomes evident that the slower moving types, creep and solifluction, deserve a much more important position than they have been accorded in the past" (1938, p. 16). He consequently devoted much of his discussion to these movements and enumerated a number of physical or cultural features as supporting evidence. Field work in the Georgia Piedmont (Fig. 1) fails to disclose these features or suggests that they may have been erroneously interpreted. It is, therefore, with these supporting evidences of soil creep that this paper takes exception, and not with Sharpe's general classification or the importance of gravitative movements as a geomorphic process.

Soil creep is defined by Sharpe as the "slow downslope movement of superficial soil or rock debris, usually imperceptible except to observations of long duration" (1938, p. 21). While evidences of creep are said to exist on practically all soil-covered slopes, it is generally accepted that downslope progress of the soil is continuous, although successions of minute movements are partially responsible, and content of water or ice may range from zero to saturation.

Suggested indications of soil and rock creep are: (1) large joint blocks moved from their place of origin; (2) trees with trunks curved concave upslope; (3) downslope bending and drag of bedded rock and weathered veins; (4) displaced posts, monuments, roads, and railroads; and (5) lines of angular to subangular rock fragments designated *stone lines* which occur at the approximate base of the creeping soil. The lines of rock fragments called *stone lines* (Pl. 1, figs. 1, 2, 3) are emphasized as a very common evidence of creep, particularly in the Southern Piedmont and Blue Ridge provinces where they are found "in almost every sizable soil cut seen in several months field work" (Sharpe, 1938, p. 24). In order to use *stone lines* as supporting evidence of mass wastage in the Piedmont, Sharpe and others have introduced a genetic process in which soil creep is necessary.

Discussing the process Sharpe says,

> "the *stone-line* itself is composed of fragments which have become detached from dikes or other resistant layers and have been drawn along at the base of the creeping soil, which have worked their way downward in the soil profile because of the more rapid movement of the surface layers, or which have been concentrated toward the base owing to the more rapid disintegration near the soil surface" (1938, p. 24–25).

Subsequent papers by Ireland *et al.* (1939) and Eargle (1940) also advance this idea for the development of the stone layers. Ireland (1939, p. 22) states that the abundance of rock fragments which form the stone line at the base of the "B" horizon . . . "appear to result largely from the fact that the rock fragments enter the creeping mass at the base and tend to stay there." Figure 2 illustrates the formation of stone lines according to Ireland, Sharpe, and Eargle.

A process dependent upon rock fragments that "enter the creeping mass at the base . . ." is subject to criticism. A quartz vein is static, and once its projection into the creeping soil is obliterated and carried away by the movement

it can release no rock fragments except with further decomposition of the enclosing bedrock. The relationship between a quartz vein and the

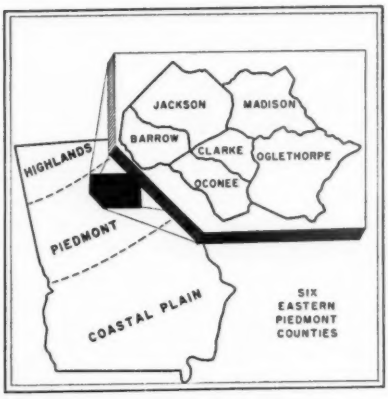

FIGURE 1.—MAP OF GEORGIA SHOWING THE SIX PIEDMONT COUNTIES IN WHICH MOST OF THE OBSERVATIONS WERE MADE

the necessary movements each particle torn from a vein discordant with the surface would have to undergo to produce an alignment of fragments by creeping soil. It seems doubtful that the pressure differential through this small vertical distance, even granting a marked variance in velocity, would be sufficient to motivate such a pronounced downward component. The increased compaction of materials beneath the fragments, the greater friction, and the necessity that material be displaced would lessen the effectiveness of the downward force. Furthermore, the initial suggestion of an undisturbed quartz vein projecting into creeping soil is basically incorrect. An undisturbed vein in weathered material must indicate an immobile condition, implying a nearly horizontal surface. Soil movements initiated by steepening the slope would probably produce fragmentation of the vein and the redistribution of the particles downslope (Fig. 4). It is conceivable that a vein lying about parallel to the moving soil layer or

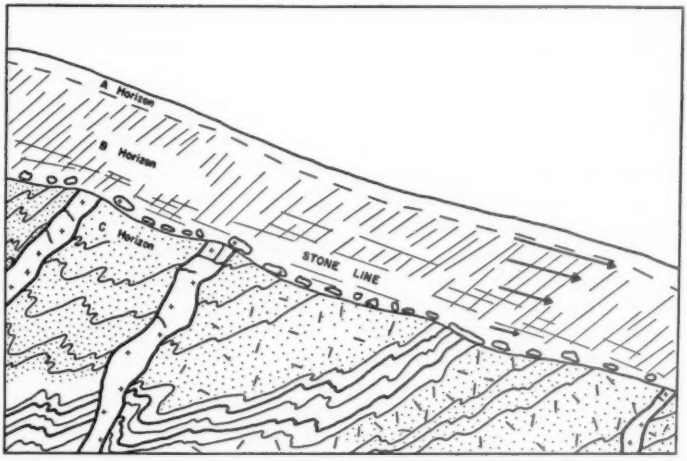

FIGURE 2.—SOIL CREEP, AND THE METHOD BY WHICH STONE LINES ARE BELIEVED TO BE FORMED
After Ireland, Sharpe, and Eargle

overlying moving soil is not analogous to the confluence of two moving media. The rock particles are an integral part of the moving mass, and, even disregarding the possibility of frost action, a downward movement resulting in horizontal alignment of the particles would depend upon shear stresses produced by varying velocities (Fig. 3A). Figure 3B illustrates intercepting it at very low angles might release fragments in a planar distribution. However, such horizontal veins are scarce and hardly can account for as common a feature as stone lines.

Finally, although Sharpe neglects frost as a factor in the formation of stone lines, Taber (1929, p. 452) has demonstrated experimentally that vertical pressures approaching 15 tons per

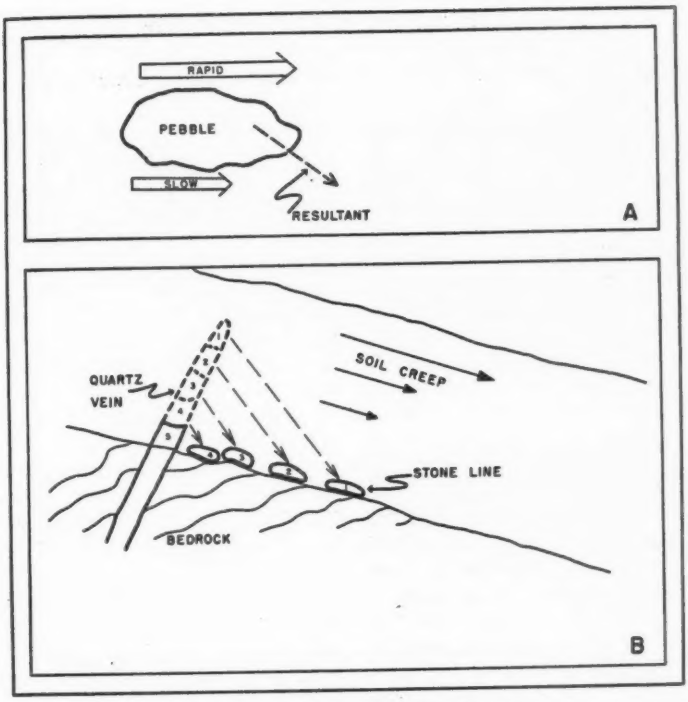

FIGURE 3.—MOVEMENT BY SOIL CREEP
A. Forces resulting from differential movement of creeping soil
B. Movements necessary to produce stone line by soil creep (no frost action)

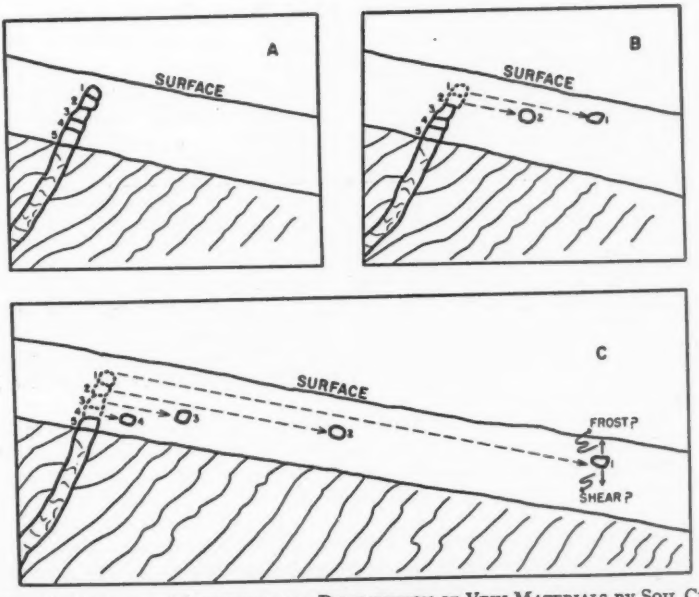

FIGURE 4.—PROBABLE MOVEMENTS AND DISTRIBUTION OF VEIN MATERIALS BY SOIL CREEP
No shear or frost action

square foot may be produced in the process of frost heaving. Pressures of this magnitude would appear more than sufficient to overcome the downward component of forces resulting from differential rates of movement within creeping soil; and, in view of the very slow movement of soils, even minor frost action should be capable of counter-balancing subsidence of the fragments. Although frost action may be negligible in the Georgia Piedmont, a supposedly similar feature has been described in areas of active frost heaving (Kerr, 1881, p. 353). Considering the above objections, residual fragments in a moving soil would probably produce a patternless mixture or surface float and not a planar distribution of the fragments at depth.

Since correlation between stone lines and slow mass wastage is doubtful, the presence of soil creep in the Georgia Piedmont must rest upon the other visible indications cited by Sharpe.

Numerous vertical cuts with pronounced surface slope were examined, and only a few exhibited even the vaguest suggestion of lineation mentioned in discussions of contemporaneous soil creep. On the contrary, predominantly vertical patterns of light lines appear quite regularly in the upper portions of the soil zone (Pl. 1, fig. 4). These lines, apparently resulting from the leaching action of organic acids along the margins of tension cracks, show no bending by surficial movements. The vertical attitude of the light pattern immediately beneath the surface is significant as the most rapid soil movements should occur in this uppermost zone.

Downslope bending and drag of bedded rocks are generally considered definite evidence of gravitative movements, and photographs of this striking feature are often included in texts. Many exposures have been examined in the Georgia Piedmont, however, in which foliation and bedding intersect the overlying soil at high angles opposed to the surface slope, and in which there is no apparent indication of any bending or drag. On the contrary, common and probably significant are steeply dipping veins (Pl. 1, fig. 5) which lie undisturbed in weathered soil mantling the slopes and crests of hills. It would be reasonable to expect that these thin fractured veins, which may extend across many feet of residual soil from the bedrock to the surface, would show some degree of distortion or separation with even slight soil motion. Absence of soil creep in this portion of the Piedmont is further suggested by the feasibility of measuring the directional properties of saprolites. These measurements are in accord with those taken on fresh bedrock, an agreement that would be improbable were there any significant disturbance by surficial movements. Again it seems unlikely that such soft unconsolidated material would not be disturbed by even the minor movement at the base of "creeping" soil.

In addition to subsurface indications already mentioned, it has been stated that natural and cultural features on the surface also reveal soil

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PLATE 1.—SOIL CREEP IN GEORGIA PIEDMONT

FIGURE 1.—Typical stone line. Scattered fragments at the base of the cut were dislodged from the stone by surface wash. Along Lexington Road, 4 miles southeast of Athens, Georgia.

FIGURE 2.—Multiple stone line composed of unusually large fragments. Along the Southern Railroad tracks near the Clarke-Jackson county line.

FIGURE 3.—Narrow stone line at considerable depth exposed by gullying. The stone line in the center of the picture can be traced continuously the full length of the gully, a distance of more than 200 yards. Half a mile northwest of Oconee Heights.

FIGURE 4.—Vertical tension cracks and vegetation roots exhibiting no bending although the surface slope is about 10°. Kaolin deposit exposed in a large gully 1 mile west of Commerce Road north of Athens, Georgia.

FIGURE 5.—Undisturbed residual quartz vein in weathered soils. The slope at this locality dips 10° to the left. This is but one of many undisturbed quartz veins. Along College Station Road, 1 mile south of Athens, Georgia.

FIGURE 6.—Large trees growing undisturbed on extreme slope. These trees have been growing on a 30-degree slope in unconsolidated fill for more than 60 years. Their straight trunks deny any subsurface soil movements. Off the Commerce Road, 1 mile north of Athens, Georgia.

FIGURE 1

FIGURE 2

FIGURE 3

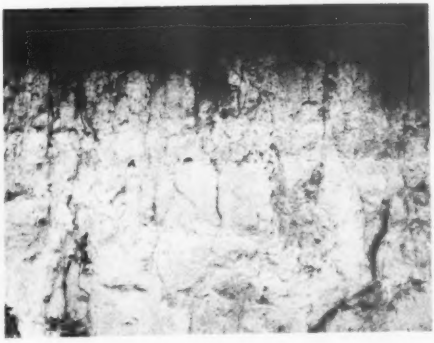

FIGURE 4

FIGURE 5

FIGURE 6

SOIL CREEP IN GEORGIA PIEDMONT

creep by their dislocation. Bending of trees, tilting of tombstones, misalignment of fence posts, displacement of roads and railroads are all used as obvious manifestations of slow gravitative soil movements. Granting that such features are supporting evidences of soil creep, although in each instance an alternate plausible displacement process can be offered, extended observations in the Georgia Piedmont reveal a definite paucity of such phenomena. Undoubtedly pertinent is the consistent lack of evidences of soil creep (Pl. 1, fig. 6) in sites where conditions should be ideal for its initiation and continuation if the process is as widespread as has been suggested.

With some doubt cast upon stone lines as evidence of soil creep, and the inability to detect other corroborating features, its importance as a contemporaneous process in the East Georgia Piedmont is questionable. Although the writers do not doubt the actuality of soil creep and the significance of other methods by which mass wastage of soil is accomplished, it is suggested that the apparent absence of soil creep in the Georgia Piedmont may be genetically significant by implying required soil texture, moisture, or temperature factors which are not present in this locality.

References Cited

Eargle, D. H., 1940, The relations of soils and surfaces in the South Carolina Piedmont: Science, v. 91, p. 337–338

Ireland, H. A., Sharpe, C. F. S., and Eargle, D. H., 1939, The principles of gully erosion in the Piedmont of South Carolina: U. S. Dept. Agriculture, Tech. Bull. 633, 142 p.

Kerr, W. C., 1881, On the action of frost in the arrangement of superficial earthy material: Am. Jour. Sci., ser. 3, v. 21, p. 345–358

Sharpe, C. F. S., 1938, Landslides and related phenomena: Columbia Univ. Press, 137 p.

Taber, S., 1929, Frost heaving: Jour. Geology, v. 37, p. 428–461

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NEW METHOD FOR TRANSFERRING DETAIL FROM AIR PHOTOGRAPHS TO A BASE MAP

BY DONALD W. LOVEJOY

During recent field work in Nevada the writer faced the problem of transferring detail from air photographs on the scale of 1:22,000 to a base map on the scale of 1:24,000. A triangulation net of horizontal control points in the field was located on the base map by plane-table intersection; the control points were selected so that they were easy to locate on the photographs and distributed so that at least three fell on each print. To transfer detail from the air photographs to the base map, a sheet of transparent acetate was mounted with drafting tape on a piece of window glass about 2 inches longer and 2 inches wider than the photograph. The glass was placed on the base map, and the control points which fell on the first photograph were traced onto the acetate with ink. The photograph was placed on the desk top and the piece of glass supported above it by means of books or any other available support; with one eye closed and the other above the center of the photograph, the writer shifted the position and tilt of the glass horizontally and vertically, until the control points on the acetate appeared superimposed on the control points on the photograph (Fig. 1). Detail on the photograph could then be traced onto the acetate with ink. The glass and acetate were next placed beneath the base map, and the detail was traced from the acetate onto the base map. The ink was wiped off the acetate with a damp cloth, and the same procedure was followed with the remaining photographs.

This method enabled the writer to transfer detail from air photographs to a base map on a smaller scale without special equipment or laborious geometric constructions. If the scale of the base map had been larger than that of the photographs, the writer would have enlarged the photographs prior to transferring the detail or introduced special equipment to magnify the photograph as the detail was transferred. The unique feature of this method is that detail is transferred from the photograph onto the base map by looking through the base map; the concept is believed to be a new contribution to photomapping and photogeology.

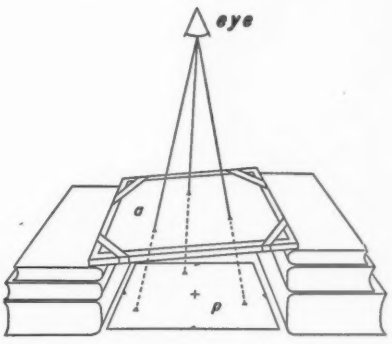
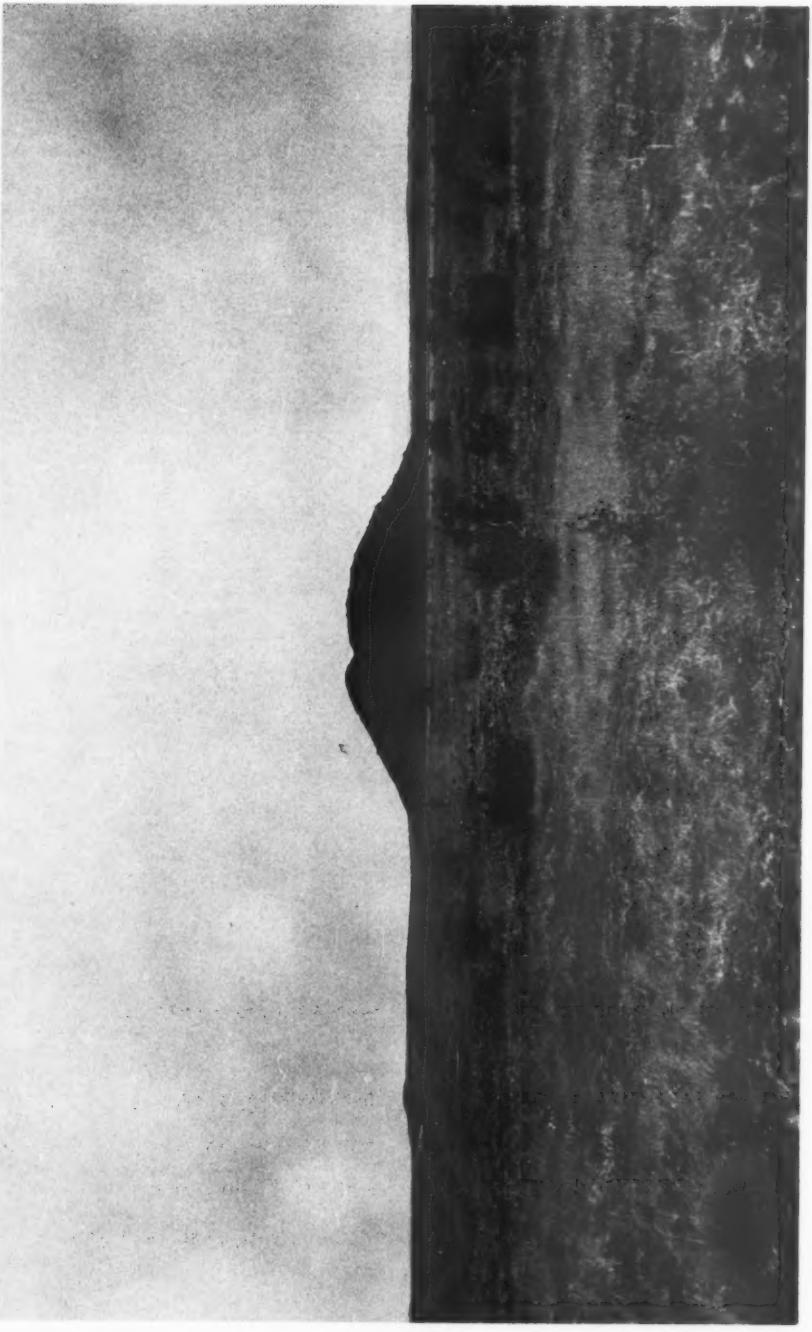


FIGURE 1.—TRANSFERRING OF DETAIL FROM AIR PHOTOGRAPH

The eye is above the picture center so that control points on the acetate (a) appear superimposed on the control points on the photograph (p).

COLUMBIA UNIVERSITY, NEW YORK, N. Y.
MANUSCRIPT RECEIVED BY THE SECRETARY OF THE SOCIETY, MARCH 23, 1956



GROUND VIEW SHOWING PINGO AND SURROUNDING TERRAIN

BULLETIN OF THE GEOLOGICAL SOCIETY OF AMERICA
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PINGO IN THE MACKENZIE DELTA, NORTHWEST TERRITORIES, CANADA

BY J. A. PIHLAINEN, R. J. E. BROWN, AND R. F. LEGGET

Pingoes are conical hills peculiar to arctic regions. They have been described by a number of investigators and travellers in the arctic, notably Porsild (1938). In the area immediately east of the Delta of the Mackenzie River in the Northwest Territories of Canada pingoes are seen in what appears to be all stages of development, from a circular or oval hillock with a pronounced crack along the major axis to a conical hill in which the original longitudinal crack has developed into a crater, and then to what appears to be the last stage, that of the crater enlarging until the pingo looks like a doughnut or a ringlike ridge above the surrounding country. Although this note records the results of a study of only one pingo, it is believed that this is the first such detailed investigation in North America. It is hoped that the information will contribute to the understanding of, and stimulate comment on, these arctic phenomena.

The opportunity to investigate the pingo arose in the course of a survey in 1954 of the lower Mackenzie Valley to determine possible new sites in the relocation of Aklavik. The survey team was headed by C. L. Merrill of the Canadian Department of Northern Affairs and National Resources; members of the Permafrost Section of the Division of Building Research of the National Research Council (Canada) constituted a part of this team and were responsible for investigating soil conditions at possible sites. Soil investigation included drilling into the perennially frozen soil (permafrost) with a specially designed portable drill rig. It was possible to penetrate to depths of 50 feet in permafrost and, more important, to obtain cores of the frozen material to this depth.

A general reconnaissance survey was made of the pingo, and two borings were drilled, one in the crater and one at the outer edge.

General Description

The pingo is located at approximately 69° 02′ N. Lat. and 134° 25′ W. Long. (Figs. 1–4; Pl. 1). It is a conical hill approximately 560 feet in diameter with its crater appreciably off center. Its summit is almost exactly 100 feet above the level of the surrounding ground. The sides of the pingo are relatively smooth, but the rim of the crater is broken so as to form four separate lips separated by sharply sloped guts approximately at right angles to one another. Dimensions of the guts are as follows:

	Length (Feet)	*Width at bottom* (Feet)	*Slope*
North Gut	65	3	35–45°
South Gut	45	3	35–40°
East Gut	300	6	45°
West Gut	30	3	35–45°

The crater is roughly circular with a diameter of about 30 feet. A pool of water in the crater at the time of the investigation was 10 feet wide, 30 feet long, 18 inches deep, and drained through the south gut. The tops of the four lips were 14, 18, 25, and 32 feet above the crater water elevation.

"Satellite" Pingoes

There are two smaller "satellite" pingoes adjacent to the main pingo. The closer one is approximately 300 feet southeast of the main pingo. It is about 60 feet in diameter and 6 feet high. A main crack or trench averaging 2 feet wide and 3 feet deep runs north-south across its diameter. Other smaller fissures oriented at random were also noted. Vegetation consisted of scrub willow up to 2 feet high with occasional willow up to 4 feet high.

The second satellite pingo is approximately 1000 feet west of the main pingo. It is 15 feet high and averages 100 feet in diameter. Vegetal

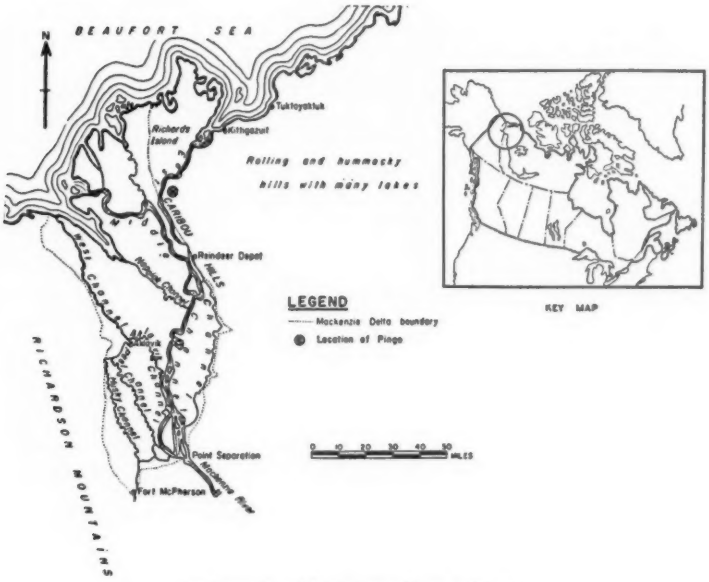

FIGURE 1.—MACKENZIE RIVER DELTA

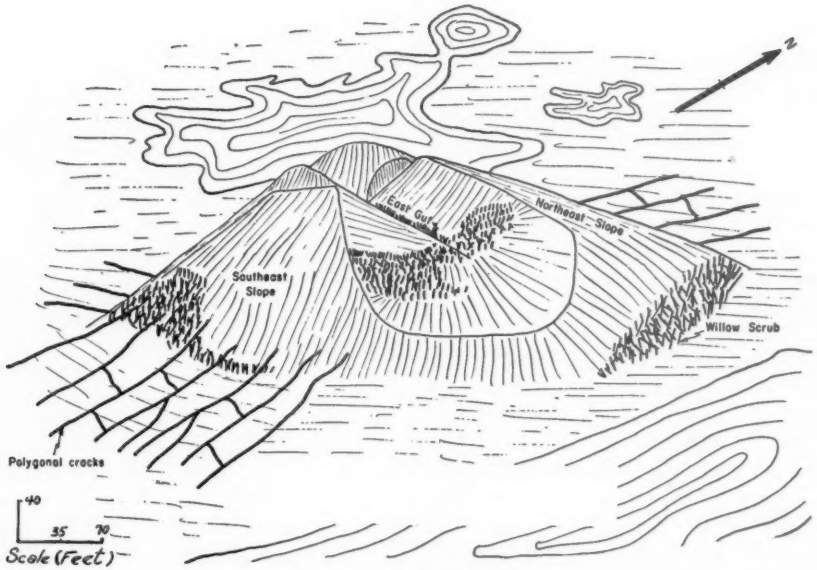

FIGURE 2.—SKETCH OF PINGO

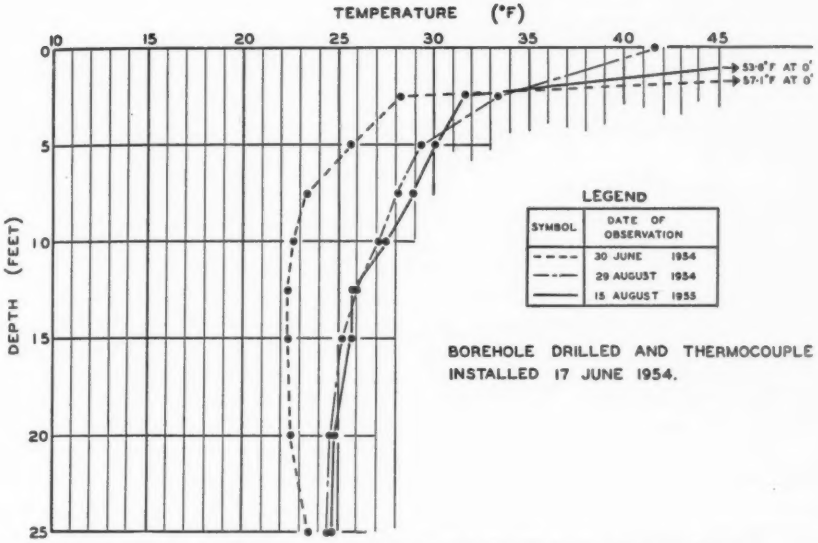

FIGURE 3.—PINGO SOIL TEMPERATURE VERSUS DEPTH BELOW CRATER SURFACE

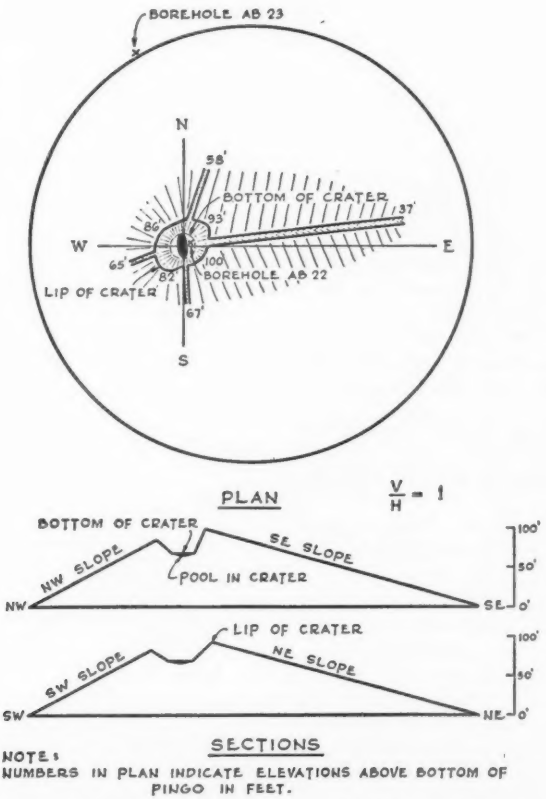

FIGURE 4.—PLAN AND CROSS SECTION OF PINGO

cover is dense willow. There is a deep fissure in its northwest face approximately 4 feet wide and 3 feet deep.

Vegetation

The area surrounding the pingo had a vegetal cover of moss and low grasses with which were associated some green sphagnum moss and wintergreen. In contrast to this the upper part of the pingo, and particularly all four guts, were covered with dense thicket up to 12 feet high, mainly willow, but with some alder as well on the east side. The thicket growth was most dense on the southwest slope and thinnest on the north. Parts of all slopes especially at the tops were devoid of vegetation.

Drilling Records

One test hole was sunk to a depth of 32 feet from a point in the crater adjacent to the pool. Good cores gave the following profile:

0 to 3 feet 8 inches. Blackish-gray silt, fine sand, with much decomposed organic material

3 feet 8 inches to 17 feet. Ice, milky because of numerous small air bubbles; ice had no apparent structure except for a few portions which were broken and lenslike

17 feet to 32 feet 3 inches. Ice, milky because of small air bubbles; core composed of many discs from a quarter of an inch to 1 inch thick

The second hole was then sunk at the lower edge of the pingo (Fig. 4), and the profile given by the record of this hole was as follows:

0 to 9 feet. Blackish-gray silt with some clay and a little fine sand with much decomposed organic material and some fibrous undecomposed organic material

9 to 13 feet. Brownish sandy silt with streaks of peat and small white shells and occasional small yellow seeds

15 feet to 17 feet 8 inches. Blackish-gray sandy silt with some clay and streaks of peat with rounded and angular pebbles up to 1 inch in diameter

17 feet 8 inches to 27 feet 8 inches. Light-gray fine sand with large subrounded but angular stones up to 1 inch in diameter. Very slow rate of drill penetration from 23 feet to 27 feet 8 inches.

Ground Temperatures

In an attempt to determine temperature conditions within the pingo a "thermocouple string" 37 feet long was inserted in the crater drill hole. It provided thermocouples to measure ground temperatures at depths of 0 feet, 2 feet 6 inches, 5 feet, 7 feet 6 inches, 10 feet, 12 feet 6 inches, 15 feet, 20 feet, and 25 feet. Readings were taken by means of a portable potentiometer.

A record of the soil-temperature observations below the surface of the pingo crater is shown in Figure 3. Soil temperatures recorded on June 30, 1954, 13 days after the hole was drilled, decrease from 28.3°F. at a depth of 2 feet 6 inches to 23.6°F. at a depth of 25 feet. Minimum recorded temperatures (22.4°F.) were found at the 12-foot-6-inch and 15-foot depths. Soil-temperature observations on August 29, 1954, revealed considerably higher temperatures decreasing from 33.4°F. at 2 feet 6 inches to 24.6°F. at the 25-foot depth. Approximately a year later (August 15, 1955) the pingo was revisited, and a set of temperature readings was taken. These temperatures were similar to those taken on August 29, 1954 (Fig. 4).

References Cited

Gussow, W. C., 1954, Piercement domes in Canadian Arctic: Am. Assoc. Petroleum Geologists Bull. v. 38, p. 816–853

Leffingwell, E. deK., 1919, The Canning River Region, Northern Alaska: U. S. Geol. Survey Prof. Paper 109, p. 150–155

Porsild, A. E., 1938, Earth mounds in unglaciated Arctic Northwestern America: Geog. Rev., v. 28, p. 46–58

Richards, H. G., 1950, Postglacial marine submergence of Arctic North America with special reference to the Mackenzie Delta: Am. Philos. Soc. Proc., v. 94, p. 31–37

Sager, R. C., 1951, Aerial analysis of permanently frozen ground: Photogrammetric Engineering, v. 17, p. 551–571

Sharp, R. P., 1942, Ground ice mounds in the tundra: Geog. Rev., v. 32, p. 417–423

Stager, J. K., 1956, Progress report of the analysis of the characteristics and distribution of pingos east of the Mackenzie Delta: Canadian Geographer, no. 7, p. 13–20

National Research Council, Ottawa, Canada
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BULLETIN OF THE GEOLOGICAL SOCIETY OF AMERICA
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EFFECT OF RADIATIVE TRANSFER ON TEMPERATURES IN THE EARTH

BY SYDNEY P. CLARK, JR.

Calculations of the temperature in the interior of the earth have previously been made under the tacit assumption that the material of the mantle is opaque to electromagnetic radiation. Such an assumption is implausible if the mantle is composed of silicates or oxides, and the purpose of this note is to illustrate the importance of radiative transfer in the earth for a few simple cases.

TABLE 1.—TEMPERATURES AT DIFFERENT RADII FOR THREE VALUES OF THE EXTINCTION COEFFICIENT
(Temperatures in °C.)

ρ	1.00	0.98	0.96	0.94	0.92	0.90	0.70	0.50	0.00
$\epsilon = \infty$	0	2400	4200	5500	6400	6900	8000	8800	9600
$\epsilon = 100\ cm^{-1}$	0	2100	3000	3500	3800	3900	4200	4400	4500
$\epsilon = 10\ cm^{-1}$	0	1800	2100	2300	2400	2500	2600	2700	2800

The importance of the contribution of radiation to heat transfer in nonopaque media is demonstrated by recent work with glasses and ceramics. The apparent thermal conductivity of hot glass is as much as 50 times as great as that of the same material at room temperature; the discrepancy has been found to be due to thermal radiation passing through the glass (Kellett, 1952). McQuarrie (1954) found that the apparent thermal conductivity of oxide ceramics increases markedly between 1500°C. and 1800°C. Ordinary conduction by lattice vibrations is incapable of accounting for this effect, and McQuarrie attributed it to radiative transfer.

The theory of the contribution of radiation to the transfer of heat is given by Van der Held (1952). For the case of steady temperatures in a medium with heat generation, the temperature is given by

$$\left(K + \frac{16n^2\sigma T^3}{3\epsilon}\right)\nabla^2 T = -A \qquad (1)$$

where K is the ordinary thermal conductivity (assumed to be constant), n is the index of refraction, σ is the Stefan-Boltzmann constant, T is the absolute temperature, ϵ is the extinction coefficient of the material, and A is the rate of heat generation per unit volume. In this form of the theory both ϵ and n are assumed independent of temperature, pressure, and the wave length of the radiation. This simplification is sufficient for present purposes.

Equation (1) was integrated for temperatures dependent only on ρ, the radial co-ordinate of a spherical co-ordinate system normalized to the radius of the earth. K was taken to be 0.006 cal./cm.sec.°C., and n^2 was taken equal to 3 (about the value for olivine or MgO in the visible part of the spectrum). The quantity A was set equal to 0.02×10^{-13} cal./cm.^{3} sec. for ρ greater than 0.9 and 0.003×10^{-13} cal./cm.^{3}sec. for ρ less than 0.9. These values were chosen to give a surface heat flow of 1.2×10^{-6} cal./cm.^{2}sec. Calculations were made for $\epsilon = \infty$, 100, and 10 cm^{-1}. The optical and thermal properties were assumed independent of ρ; in particular the core was neglected.

Resulting temperatures at several depths are given in Table 1. The calculations were made by assuming that the effective conductivity,

$$C = K + \frac{16n^2\sigma T^3}{3\epsilon}, \qquad (2)$$

was constant over short intervals of depth. Thus the nonlinear equation (1) was replaced by a system of approximating linear equations. Until better values of C are available, a more

accurate, but far more tedious numerical integration is unwarranted.

The effective conductivity deep in the earth is 0.03 cal./cm.sec.°C. for $\epsilon = 100$ cm.$^{-1}$, and 0.06 cal./cm.sec.°C. for $\epsilon = 10$ cm.$^{-1}$. Thus the effective conductivity is 5 or 10 times as large as the ordinary conductivity for the above values of the various constants.

Near the surface of the earth, the temperature gradient is steep because the temperatures are low, and radiative transfer is unimportant. At a depth of a few hundred kilometers, the effective conductivity increases markedly, and the thermal gradient is correspondingly reduced. These features of the curve illustrate the "safety-valve" effect, arising from the monotonic increase of the effective conductivity with temperature. An increase in temperature is partially offset by an increase in effective conductivity. Because of this, the temperature deep in the mantle may be relatively insensitive to different assumptions about the value of the extinction coefficient and the distribution of radioactive heat sources.

A detailed analysis of the magnitude of the extinction coefficient and its dependence on temperature, pressure, and wave length for materials likely to exist in the mantle must be made before the importance of radiative transfer in the earth can be properly assessed. Preliminary results show that the extinction coefficients of basic rocks are between 20 and 40 cm.$^{-1}$ at wave lengths of about 0.8 microns at room temperature. The results shown in Table 1 indicate that, if ϵ is as low as 100 cm.$^{-1}$, radiative transfer will drastically modify the temperature at depth.

References Cited

Kellett, B. S., 1952, Transmission of radiation through glass in tank furnaces: Jour. Soc. Glass Technology, v. 36T, p. 115–123

McQuarrie, Malcolm, 1954, Thermal conductivity: VII, Analysis of variation of conductivity with temperature for Al_2O_3, BeO, and MgO: Jour. Am. Ceramic Soc., v. 37, p. 91–95

Van der Held, E. F. M., 1952, The contribution of radiation to the conduction of heat: Applied Sci. Research, A, v. 3, p. 237–249

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BULLETIN OF THE GEOLOGICAL SOCIETY OF AMERICA
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REGARDING THE EARLY HISTORY OF THE EARTH'S ATMOSPHERE

BY HAROLD C. UREY

From a calculation which the writer believes is not conclusive on this question, Rubey (1955) concludes that the early atmosphere of the earth could not have been a reducing one. If we assume that the quantities of the elements H, O, C, and N now in the surface regions of the earth were present on the earth at any time in the past, the condition of thermodynamic equilibrium will give the results secured by Rubey. However, if there were a larger proportion of hydrogen, then results would be quite different. It is difficult to solve the equations proceeding in this way, and the writer proceeds differently.

We assume a slight excess of hydrogen, let us say, enough to give a partial pressure of 10^{-3} atmosphere, of elementary hydrogen as was assumed previously by the writer (Urey, 1952a). The equilibrium constant for the reaction,

$$C \text{ (graphite)} + 2\ H_2 \text{ (gas)} = CH_4 \text{ (gas) is } 10^{8.9},$$

and thus

$$p_{CH_4} = p_{H_2}^2 \times 10^{8.9}.$$

If p_{H_2} is 10^{-3}, then p_{CH_4} at equilibrium is 794 atmospheres. This is a 100 times greater pressure than would exist with all surface carbon converted to methane. Hence no graphite need be present. The equilibrium constants for the reactions,

$$CO_2 \text{ (g)} + 4\ H_2 \text{ (g)} = CH_4 \text{ (g)} + 2\ H_2O \text{ (liquid)}$$

and

$$CO \text{ (g)} + 3\ H_2 \text{ (g)} = CH_4 \text{ (g)} + H_2O \text{ (liquid)}$$

are $10^{+22.9}$ and $10^{+26.4}$ respectively. The partial pressure of the water does not appear in these equilibrium constants and hence

$$K = \frac{p_{CH_4}}{p_{CO_2} p_{H_2}^4} = 10^{22.9} \text{ or}$$

$$p_{CO_2} = 10^{8.9} \times 10^{12} \times 10^{-22.9} = 10^{-10}$$

$$K = \frac{p_{CH_4}}{p_{CO} p_{H_2}^3} = 10^{26.4} \text{ or}$$

$$p_{CO} = 10^{8.9} \times 10^{9} \times 10^{-26.4} = 10^{-16.5}$$

where the pressure of methane is taken as 7.94 or $10^{0.9}$ atmospheres, and we see that the partial pressures of CO_2 and CO are very small. Hence carbon would exist mostly as methane, if equilibrium were obtained. The equilibrium constant for the reaction

$$N_2 \text{ (g)} + 3\ H_2 \text{ (g)} = 2\ NH_3 \text{ (g)}$$

is $10^{5.82}$. Hence if p_{H_2} is 10^{-3}, and p_{N_2} = 0.8 atm., the p_{NH_3} is $10^{-1.6}$ atm., and thus most of the nitrogen would not be present as ammonia gas. (The surface partial pressure of N_2 would not be 0.8 atm. as it is now because of admixture with large amounts of methane, but this would change numerical values only by small and unimportant factors.) However, ammonia is very soluble in water, and if oceans were present some ammonia would exist even if the partial pressure of hydrogen were 10^{-3} atm. If organic acids were present in the oceans, ammonium ion, which is more stable than ammonia, would be present. Calculation shows that, for half the nitrogen to be present in the oceans as NH_4^+, the concentration of NH_4^+ must be about 0.1 N, and that this could occur even with the water somewhat on the basic side of neutrality. Miller (1955) has shown that under the conditions postulated the nitrogen would form amino acids and amines and hence would be present partly as such compounds. This complication makes all calculations uncertain. It does seem likely that the nitrogen would be present to a large extent in such compounds even if the partial pressure of H_2 were as low as 10^{-3} or 10^{-4} atm., as can easily be seen by repeating the calculations using this lower pressure.

The calculation for carbon is incorrect, however, because Miller's experiments show that carbon would be present as carbon compounds

dissolved in water and only partly as methane. Hence, all these calculations give only indications of compounds expected.

Of course, chlorine would be present as chloride ion in combination with Na^+, Mg^{++}, Ca^{++}, Fe^{++}, etc., and not as HCl. In our laboratories this acid dissolves rapidly in water and in minutes or hours attacks rocks to form the chlorides of these basic elements.

If Rubey will assume larger amounts of hydrogen relative to carbon and oxygen he will find that his calculations will show that methane can be present. Also ammonium ion will be present. Of course, neither he nor the writer knows how much hydrogen to assume, but we also cannot say that some excess was not present at least temporarily.

As pointed out by the writer some years ago, a reducing atmosphere containing methane would be a cold atmosphere at high altitudes since methane would rise very high in the atmosphere, and since it radiates in the infrared it would cool the upper atmosphere rapidly. An atmosphere containing small amounts of hydrogen and hence much methane would probably lose the hydrogen very slowly. This argument was presented in the reference cited by Rubey (Urey, 1952a), and he presents no reasons for believing that it is incorrect.

Also, Rubey remarks that an increased concentration of CO_2 would dissolve $CaCO_3$ and not result in an increase of limestone. The writer (Urey, 1952b) has suggested that the low concentration of CO_2 in our atmosphere is due to the continual and rapid removal of CO_2 by reactions of the general kind,

$$CaSiO_3 + CO_2 = CaCO_3 + SiO_2$$

or

$$MgSiO_3 + CO_2 = MgCO_3 + SiO_2$$

or other reactions involving more complex silicates, and the writer used the first reaction to illustrate the situation. If we consider this reaction and the reaction of solutions of CO_2 with $CaCO_3$, *i.e.*,

$$CaSiO_3 + CO_2 = CaCO_3 + SiO_2,$$

$$K = \frac{1}{p_{CO_2}} \cong 10^{+8},$$

and

$$CaCO_3 + CO_2 + H_2O = Ca(HCO_3)_2 \text{ (solution)},$$

we can discuss the question raised by Rubey. If the pressure of CO_2 in the atmosphere is increased suddenly above the equilibrium value, then both reactions are displaced to the right, the second one quickly and the first very slowly as CO_2 comes into contact with the rocks through erosion processes. The second would require times of the order of thousands of years—namely, the time required for mixing of the oceans, and the first times of the order of millions of years—*i.e.*, the time required for leveling of mountains. As erosion proceeds, CO_2 would be withdrawn from the atmosphere by the first reaction, and then the second reaction would be displaced to the left, and this will proceed until equilibrium is re-established or until some other upset of equilibrium occurs. It is the writer's intention only to point out that the assumption that all the CO_2 now in the sedimentary rocks could at any time have remained in the atmosphere for more than a few million years is unrealistic, if water and land areas existed. Also, because of such equilibria, he doubts if the earth ever had much more CO_2 (*i.e.*, 10 times more) in its atmosphere than it has now.

Rubey states that he believes these considerations are in error. Perhaps his conclusion is due to a misunderstanding. At 25°C the partial pressure of CO_2 in equilibrium with $CaSiO_3$, $CaCO_3$, and SiO_2 is 10^{-8} atm. with a probable error of a factor of 10 or more. At 700°K or 427°C this equilibrium pressure is about 40 atm. Now the earth is not at one temperature. At 25°C with a partial pressure of CO_2 of about 3×10^{-4} atm. the reaction proceeds toward $CaCO_3$ and SiO_2. At only slightly elevated temperatures it proceeds toward the formation of $CaSiO_3$ and CO_2, and the gas escapes. At least much of the CO_2 coming to the surface of the earth is produced by these reactions of calcium and magnesium carbonates with silica or other silicious materials, or other acid oxides—*e.g.*, alumina. Hence the pressure of CO_2 lies between these limits, and because of the rather rapid rates with which acids attack igneous rocks it is not surprising

that the partial pressure of CO_2 is near the lower value. The writer does not believe Rubey has given any arguments that invalidate these conclusions.

References Cited

Miller, S. L., 1955, Production of some organic compounds under possible primitive earth conditions: Jour. Am. Chem. Soc., v. 77, p. 2351; 1953, Science, v. 117, p. 528

Rubey, W. W., 1955, Development of the crust of the Earth, p. 631–650 *in* Poldervaart, Arie, *Editor*, Crust of the earth: Geol. Soc. of America Special Paper 62, 762 p.

Urey, H. C., 1952a, On the early chemical history of the earth and the origin of life: Nat. Sci. Proc., v. 38, p. 354, 359; 1952b, *The Planets*, New Haven, Yale Univ. Press, p. 148 ff.

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BULLETIN OF THE GEOLOGICAL SOCIETY OF AMERICA
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ESTIMATES OF THE AVERAGE ABUNDANCE OF Cu, Mn, Pb, Ti, Ni, AND Cr IN SURFACE WATERS OF MAINE

BY KARL K. TUREKIAN AND M. DEAN KLEINKOPF

The trace-element content of fresh surface waters has received attention recently by investigators interested in the geochemical prospecting for ores. By their very nature, these studies were restricted to small regions where detailed work seemed useful. Consequently little has been reported on the trace-element content of surface waters of large areas which might be useful for general geochemical discussions and calculations.

The present paper reports analyses on 439 samples of stream and lake waters from all over the State of Maine. The work was initiated to test the feasibility of large-scale geochemical prospecting in unexplored areas (Kleinkopf, 1955, Ph.D. thesis, Columbia Univ.), but the data are applicable to basic geochemical problems.

A semiquantitative emission spectrographic method was used in the analysis of the evaporated residues from the water samples collected. The standard concentration ranges were 1, 5, 10, 50, 100, 500, 1000, 5000, and 10,000 ppm of the metal in the residue. For six of the elements studied—Ni, Ti, Cr, Pb, Mn, and Cu—all values fell within these ranges.

In a region of limited sampling where variations are small, semiquantitative data cannot give a representation of the concentration of an element in the average surface water. In the case of the Maine lakes, however, a large number of observations (439) and a large dispersion of values (Fig. 1) give the semiquantitative values the nature of a "grouping" operation. For this reason it is felt that Table 1 gives a very close approximation to the average trace-element concentration of the surface waters of Maine.

Table 2 lists the average values for each element studied calculated as above. The first column is in terms of parts per million of the residue. The second column converts this to parts per billion of the water on the basis of an average concentration of dissolved salts of 13.5 mg/liter. This low value is characteristic of Maine where the glaciated crystalline surface provides less dissolved material than other areas.

TABLE 1.—SAMPLE CALCULATION FOR THE ESTIMATIONS OF THE ABUNDANCE OF COPPER IN THE MAINE SURFACE WATERS

(1) ppm Cu	(2) No. of Cases	(1) × (2)
50	3	150
100	80	8,000
500	249	124,500
1,000	78	78,000
5,000	25	125,000
10,000	4	40,000
Total	439	375,650

Average ppm Cu $\frac{375,650}{439} = 860$ ppm.

TABLE 2.—AVERAGE ABUNDANCE OF Cu, Mn, Pb, Ti, Ni, AND Cr IN SURFACE WATERS IN MAINE

	ppm of solid residue	parts per billion of water (13.5 mg. solids/liter)
Copper	860	1.16
Manganese	300	.40
Lead	190	.26
Titanium	150	.20
Nickel	20	.02
Chromium	20	.02

Figure 1 shows that a generally symmetrical distribution is obtained with the grouping inherent in the semiquantitative method. This is comparable to the symmetry observed in some cases for the histograms of trace elements

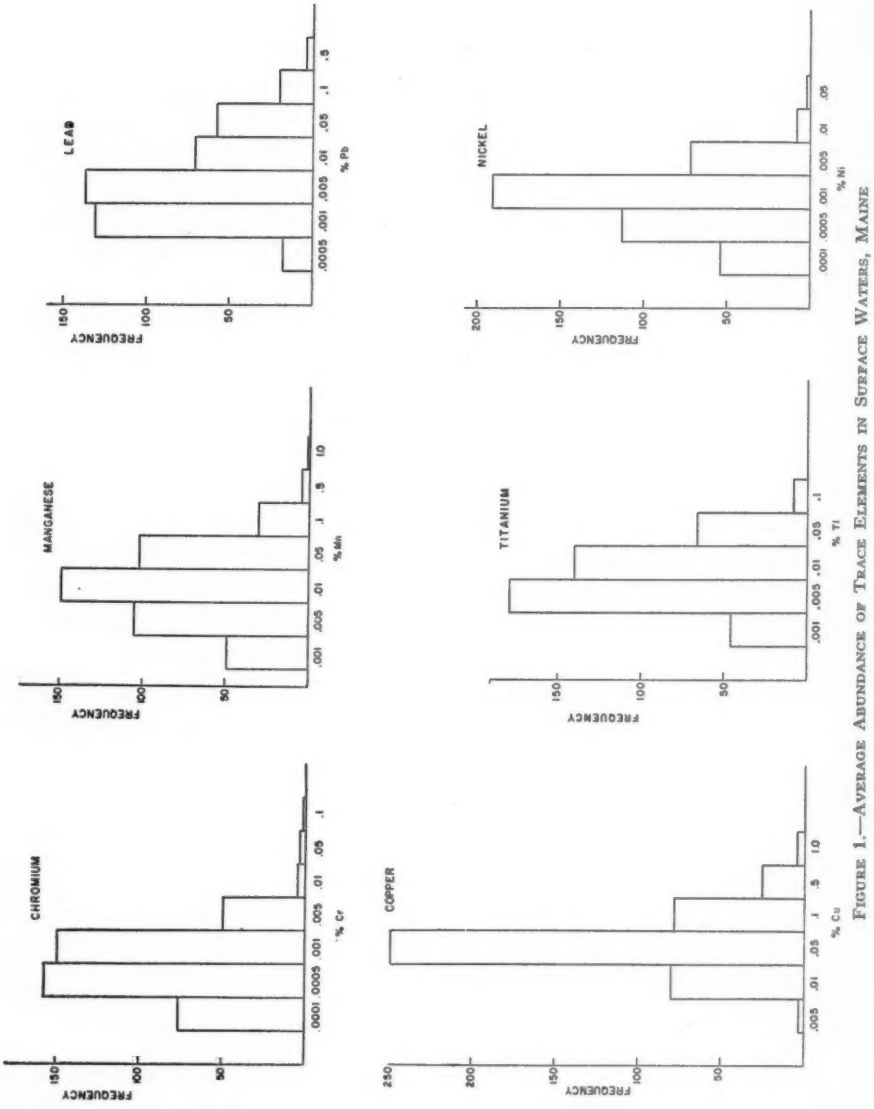

FIGURE 1.—AVERAGE ABUNDANCE OF TRACE ELEMENTS IN SURFACE WATERS, MAINE

in other natural systems where a logarithmic transformation has been made of the concentration variate (Ahrens, 1953; 1954; Turekian and Kulp, In press).

The fact that the dominant crystalline rocks in Maine are granitic is reflected generally in the low values of Ni and Cr and the high value of Pb. The rather high value of Cu may represent either a high mobility of the copper ion or the presence of copper sources other than the normal crystalline rocks.

FIGURE 1.—AVERAGE ABUNDANCE OF TRACE ELEMENTS IN SURFACE WATERS, MAINE

References Cited

Ahrens, L. H., 1953, The lognormal distribution of the elements: Geochimica et Cosmochimica Acta, v. 5, p. 49

—— 1954, The lognormal distribution of the elements (2): Geochimica et Cosmochimica Acta, v. 6, p. 121

Turekian, K. K., and Kulp, J. L., In press, The geochemistry of strontium: Geochimica et Cosmochimica Acta

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